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Spatial Variability of Two Different Soil Moisture Regimes

by

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A THESIS
SUBMITTED TO THE FACULTY OF GRADUATE STUDIES AND RESEARCH
IN PARTIAL FULFILMENT OF THE REQUIREMENTS FOR THE DEGREE
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IN
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Date. Oct. 13 .. 1984 ....

Abstract

Optimization of land management for water conservation, soil salinity and erosion control requires a through understanding of the soil water balance. The adequate design of any water balance study requires that the precision and the spatial variability of the soil parameters be known. The objectives of this study were to quantify the soil water balance components and to determine the precision and spatial variability of soil moisture and related soil properties. The study was conducted during 1983 at the Ellerslie Research Station near Edmonton, Alberta.

Soil moisture was measured with a neutron probe at 42 locations, 6.1 m apart, arranged in a six by seven grid. Fallow plots were alternated with barley plots resulting in three replicates of each. Particle size analysis, bulk density, and the moisture characteristic curve were determined at several depths at each location. Due to dry conditions in May and June the barley did not germinate until late June. Heavy rains during late June and early July resulted in near saturated conditions and in a perched water table within 1 m of the surface. Dry, warm weather for the remainder of the growing season resulted in vigorous barley growth.

Precipitation for the study period was 284 mm. The total change in soil moisture for the top 1 m of the soil profile was -17 mm and -64 mm for the fallow and barley plots respectively. Evapotranspiration, determined in part

from the Penman method and from the field capacity method, was 207 mm and 254 mm for the fallow and the barley plots respectively. Drainage was determined to be 94 mm for both fallow and barley plots. As it was not possible to separate deep drainage from upward flow, the contribution of the water table to evapotranspiration could not be evaluated.

An alternate method for evaluating drainage and evapotranspiration of the barley plots, the gradient method, yielded only a 2% difference in evapotranspiration from the field capacity method.

Discontinuous sand lenses below 60 cm resulted in highly skewed distributions of the particle size, the characteristic curve, and the moisture content during dry conditions. During the near saturated conditions of the heavy rainfalls the skewed moisture conditions normalized. This resulted in soil moisture sample size requirements of 4 to over 20 for a precision of ±5%. Semivariograms of soil moisture and related soil properties showed general spatial independence for distances from 6.1 to 37 m. Spatial dependence could exist at distances less than or greater than those measured.

An adequate sampling program for the determination of soil moisture for this site, which is approximately 0.1 ha in area, would require at least 10 to 15 access tubes arranged between 6 and 37 m apart to achieve a precision of ± 5%. For these distances systematic sampling does not offer any increase in precision over random sampling.

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To my fellow graduate students I extend my appreciation for their efforts in making my sentence here absurd and likened to the inside of a Monty Python sketch, that is greatly enjoyable. These fellow graduate students have not only helped out in innumerable ways, but have also become my friends. I would also like to not extend special thanks to the organization F.O.P. for the help they did not provide.

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1. INTRODUCTION

Quantification of the soil water balance components (precipitation, drainage, and evapotranspiration) is vital to the understanding and management of agricultural and forested lands. Knowledge of drainage is necessary for proper irrigation scheduling, salinity control, and prevention of groundwater contamination. The study of potential and actual evapotranspiration aids in the development of better management practices for optimizing crop growth.

Soil is a particulate matrix of mineral and organic matter that contains air and water in the voids. The relative proportions of these constituents change with distance due to the geologic history, groundwater, slope, vegetation, and climatic influences.

Studies of the heterogeneity of soil-water properties indicate that appreciable ranges in variation may be encountered. Generally to obtain a statistically reliable value for soil moisture and bulk density only several samples are required, whereas for soil water flux properties (e.g. changes in soil moisture, hydraulic conductivity, infiltration, and diffusion) often over one hundred samples are required to achieve the same degree of precision.

Soil is also a continuous body, in which the relationship between neighbouring locations increases as distance between these locations decreases. Because soil properties vary, an estimate of the degree of representation of the total population is needed. Proper representation of soil variability should include not only an estimate of the variability, but also an estimate of how the variability changes with distance. Knowledge of the spatial variability of soil water is valuable for the design of water balance studies. With estimates of population representation and spatial dependence, the establishment of proper sample intervals can be established.

The purpose of this study was to determine the degree of variability that is encountered in soil water changes due to precipitation, drainage, and evapotranspiration. The objectives of this study were:

1. To quantify the soil water balance components: precipitation, drainage, evapotranspiration, ground water contribution, and changes in soil storage for barley and fallow plots.
2. To determine the population distribution and spatial variability of soil moisture storage changes, moisture content, texture, bulk density, and moisture characteristic curve.

2. SITE DESCRIPTION AND EXPERIMENTAL DESIGN

2.1 SITE DESCRIPTION

2.1.1 LOCATION

The study site was located about 15 km south of the Edmonton city center at the Ellerslie Agricultural Research Station (NE 1/4 Sec 24, Tp 51, R 25, W 4; lat. 53° 25' N, long. 113° 33' W). A fully equipped meteorological station is situated at the station, adjacent to the study site.

2.1.2 CLIMATE

The climate of the Edmonton area is cool continental characterized by relatively warm summers and cold winters (Bowser *et al.*, 1962). The average annual temperature is 3°C, with January the coldest month at -14°C and July the warmest month at 16°C (Table 1). The average frost free period is greater than 100 days (Crown and Greenlee, 1978).

The climate is between dry and moist subhumid with a mean annual precipitation of 338 mm. Sixty percent of the annual precipitation falls in the period May-August, with July having the greatest amount (Table 1). Rainfall accounts for 70% of the precipitation, with the rest occurring as snow (Verma, 1968). The rainfall during the growing season has been described as low in intensity and well distributed (Toogood, 1963). Laycock (1967) reported that the moisture deficit for the growing season is 5 to 21 cm assuming a 10

TABLE 1. SUMMARY OF CLIMATIC DATA RECORDED AT EDMONTON INTERNATIONAL AIRPORT FROM 1941-1970
(adapted from Crown and Greenlee, 1978)

	MONTH							YEAR
	A	M	J	J	A	S	O	
Temperature (°C)								
Mean daily	3	10	13	16	14	10	4	1.4
Mean daily max.	9	17	20	23	21	17	11	7.8
Mean daily min.	-3	2	6	9	7	3	-3	-5.1
Extreme max.	26	30	34	35	33	34	27	35
Extreme min.	-19	-18	-6	0	-1	-8	-19	-45
Precipitation								
Mean rainfall (mm)	9	32	76	99	62	41	9	338
Greatest rain 24 hrs(mm)	12	21	66	49	34	60	7	66
No. of days of rain	3	9	12	13	11	9	5	66

cm soil water capacity. Cohen (1980) reported a mean annual soil moisture deficit of approximately 10 cm and mean annual potential evapotranspiration of approximately 50 cm.

Wind velocity for the year averages 16 km h^{-1} with the dominant direction being from the NW. Sunshine averages 2,175 hours during the year (Bowser *et al.*, 1962).

2.1.3 VEGETATION

The Ellerslie Agricultural Research Station is located in the forest-grassland transition. Undisturbed vegetation of the immediate area is mostly a balsam poplar forest. Balsam poplar (*Populus balsamifera*) is the dominant forest species in the canopy with a strong admixture of aspen poplar (*Populus tremuloides*) on the upper slopes and white spruce (*Picea glauca*) on the lower slopes (Pawluk and Dudas, 1982).

The study plots are located on a cultivated portion of the farm. A large portion of the area is under pasture and mixed grain cultivation, with barley being the popular grain crop (Cohen, 1980).

2.1.4 SURFICIAL GEOLOGY

The Ellerslie Agricultural Research Station is on a gently rolling to rolling morainal plain with slopes rarely exceeding 2% except at river and creek channels (Crown and Greenlee, 1978; Cohen, 1980). The study plots are located on a local rise (for several km) with gentle slopes (less than

2%) in all directions, except to the south where the land is nearly level.

The Laurentide ice sheet advanced over the entire area in a general southerly direction depositing a mantle of till of variable thickness (Crown and Greenlee, 1978). Much of Edmonton area, including the Ellerslie Agricultural Research Station, is located on the Edmonton Glacial Lake Plain. The deposits of the Proglacial Lake Edmonton covered the till with a blanket of fine-textured glaciolacustrine sediments up to 16 m thick. Pawluk and Dudas (1982), in a study, about 500 m to the north of the plots, (Figure 1) found that sorted drift was frequently interbedded with lacustrine materials and formed the surficial materials of the surrounding site. Bowser *et al.* (1962) found that the soil series that occurs in this area (Malmo) frequently developed on slightly saline lacustrine material 1 to 1.7 m thick, separated from the till by a sand layer.

2.1.5 HYDROGEOLOGY

About 150 metres to the south of the plots there is a 2 to 3 metre drop in elevation to the east (Figure 1). Over a distance of approximately 40 meters on this slope several soil pits have been dug and described (Crown and Greenlee, 1978). At the bottom of this slope is a discharge area with a Rego Humic Gleysol in a depression. A water table has been recorded within 10 to 40 cm of the surface for 3 out of 4 years (Pawluk, 1981). The water table represents discharge

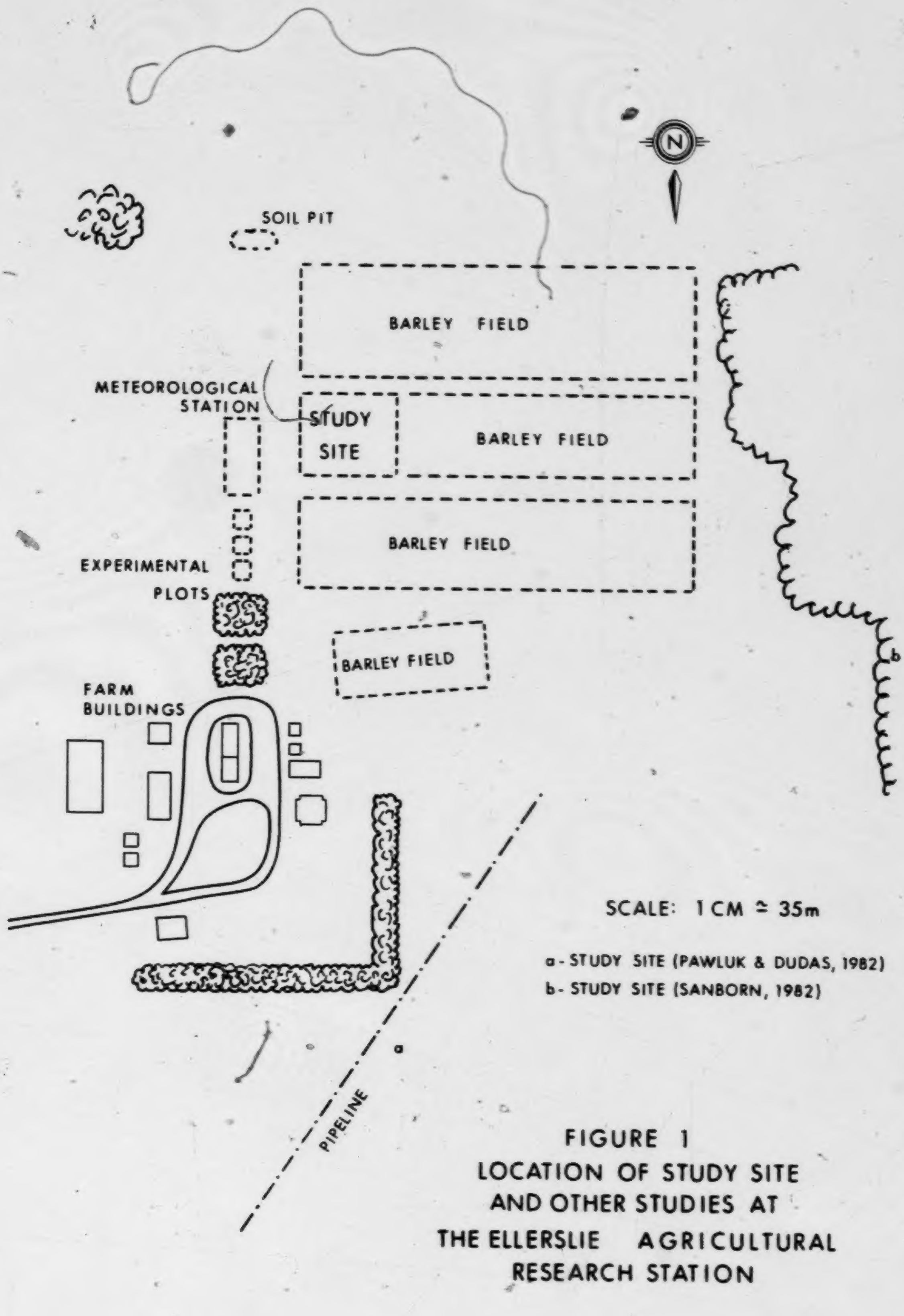

SOIL PIT
BARLEY FIELD
METEOROLOGICAL STATION
STUDY SITE
BARLEY FIELD
BARLEY FIELD
EXPERIMENTAL PLOTS
BARLEY FIELD
FARM BUILDINGS
b
a
PIPELINE
N
SCALE: 1 CM ≃ 35m
a - STUDY SITE (PAWLUK & DUDAS, 1982)
b - STUDY SITE (SANBORN, 1982)

FIGURE 1
LOCATION OF STUDY SITE
AND OTHER STUDIES AT
THE ELLERSLIE AGRICULTURAL
RESEARCH STATION

from the slope probably as the result of a sand layer redirecting flow. A soil pit located just above the toe of the slope has few fine distinct mottles occurring at 48 to 62 cm (Crown and Greenlee, 1978) probably indicative of a capillary fringe from a variable water table (Pawluk, 1981). The Gleysolic soils may be saline or carbonated (Crown and Greenlee, 1978).

Sanborn (1981) studied the piezometric surfaces at a location about 800 m to the NW of the study site (Figure 1). He found that the head was highest in the 10.4 m piezometer, while a 4.5 m piezometer was dry throughout the study period (1977 to 1978). This was interpreted as indicating upward movement of groundwater into a more permeable sand and gravel layer, discharging elsewhere in the landscape. Piezometric levels were higher during the spring and continued rising until late June. A 3.6 m water well remained dry except during May, 1979, when the water level was within 1 to 1.5 m of the surface. This, according to Sanborn (1981), indicated a perched water table as the 4.5 m piezometer remained dry.

2.1.6 SOILS

Soils in the well- to moderately-well drained positions of the Ellerslie Agricultural Research Station have been described as Eluviated Black Chernozems (Bowser *et al.*, 1968; Verma, 1968; Crown and Greenlee, 1978; Sanborn, 1981; Pawluk and Dudas, 1982). These include the Malmo Series and are found in level to gently undulating areas characteristic of the lacustrine deposits of Proglacial Lake Edmonton (Bowser et al., 1962). Gleyed Eluviated Black Chernozems and possibly Solodic Black soils are found in the imperfectly drained positions (Crown and Greenlee, 1978).

2.2 EXPERIMENTAL DESIGN

Plot design must allow for maximum flexibility in satisfying experimental objectives, often involving compromises. The following rationale was used in the experimental design (Figure 2).

1. The water balance components of summerfallow and barley plots were to be quantified and compared.
2. The determination of spatial dependence required that sampling sites have spatial coordinates. The simplest coordinate system for both layout and analysis is a grid.
3. The effects of soil variability between the fallow and barley plots had to be minimized; thus, the treatments had to be as close together as possible without edge effect. Consequently adjacent rectangular plots with alternating treatments were favoured over separate, large plots.
4. Locations of moisture measurements at least 2.5 m away from plot edges should minimize edge effect. Consideration of farm machinery size led to a final

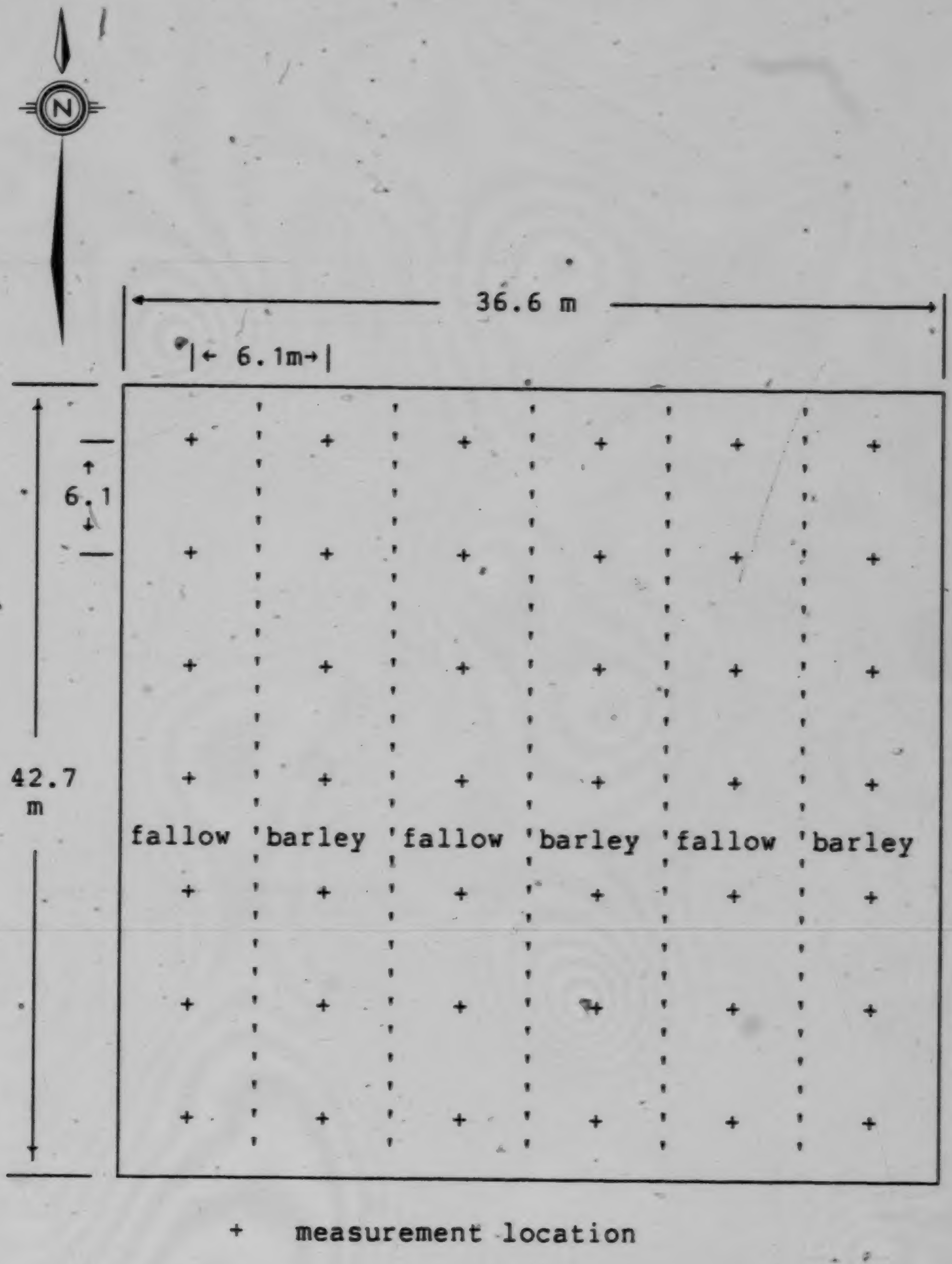

N
36.6 m
6.1m
6.1
42.7 m
fallow
barley
fallow
barley
fallow
barley
+ measurement location

Figure 2. Site layout

spacing between sample points of 6.1 m.

5. Four to ten gravimetric samples are needed to obtain an accurate estimate of soil moisture content (Hillel, 1980b). Using the neutron probe, seven sample sites arranged linearly should give sufficient accuracy and distance for spatial determinations of soil moisture content.
6. Three replicates of fallow and barley plots should aid with statistical comparisons between treatments and the establishment of spatial dependence for a two-dimensional surface.

3. SOIL WATER BALANCE

3.1 LITERATURE REVIEW

3.1.1 INTRODUCTION

Determination of a soil water balance entails quantification of the balance components. Components of water gain are; precipitation, irrigation, interflow and runoff from upslope, or from a shallow ground water table. Components of water loss are; drainage, evaporation, transpiration, and runoff or interflow downslope. Using a model of *input equals output plus change in storage*, a model for soil water balance may be formed:

$$P + I + U = E + T + D \pm Ro \pm L + \Delta S \qquad (1)$$

where, P = precipitation,
I = irrigation,
U = capillary rise,
E = evaporation,
T = transpiration,
D = drainage,
Ro = runoff,
L = interflow,
ΔS = change in soil moisture storage.

The ± symbol indicates the component can result in either water gain or loss.

Assuming a flat land surface with negligble runoff, interflow or capillary rise and with no irrigation, equation *1* may be simplified to:

$$P = Et + D + \Delta S \qquad (2)$$

where $Et = E + T$ (evapotranspiration)

3.1.2 EVAPORATION FROM BARE SOILS

3.1.2.1 Introduction

Evaporation is the transformation of liquid water to a vapour. For this to occur there must be a supply of about 2472 joule/g for vaporization (Penman, 1963). The amount of water evaporated from the soil surface is dependent upon the evaporative demand (atmospheric conditions such as wind, temperature, and radiative energy) and soil water availability.

3.1.2.2 Stages of Evaporation

Evaporation from an initially saturated soil may be divided into three successive stages (Lemon, 1956; Hillel, 1980b): an initial constant-rate stage in which the evaporation ~~rate~~ is limited only by atmospheric conditions; an intermediate falling-rate stage in which the rate rapidly declines and comes under control of soil conditions; and a slowrate stage in which water transmission occurs primarily by vapour diffusion.

The transition from the first to the second stage is generally sharp, while the second stage blends into the third stage so gradually that they cannot be separated easily. Some researchers prefer to combine the second and third stages and to recognize only two stages (Ritchie and Jordan, 1972; Hanks and Ashcroft, 1980). Jackson *et al.* (1973) found that the three stages could not be delineated under natural conditions.

During the first stage, the steepness of the soil moisture gradient is determined by the rate of loss and by the diffusion coefficient governing the relationship between the rate of flow and the moisture gradient (Lemon, 1956). Vapour flow is insignificant. Immediately after wetting the evaporation from a wet bare soil is approximately the same as that from a free water surface at the potential evapotranspiration rate (Lemon, 1956; ASCE, 1973; Stewart, 1984). Hartmann *et al.* (1980) found good agreement between bare soil evaporation and potential evapotranspiration (*PET*) estimated by the Penman equation corrected for a free water surface by multiplying by 0.8 for May to August.

The length of the first stage depends upon water availability and meteorological conditions. Water availability is dependent upon texture, with finer-textured soils having a longer first stage than coarse-textured soils (Wilcox, 1960; Hillel, 1980b). When meteorological conditions result in low evaporation rates, the initial, constant-rate stage persists longer which, according to Gardner and Hillel (1962), results in lower cumulative losses. Lemon (1956) found that liquid flow was insufficient to match evaporative demand as soil potential decreased to -30 kPa.

During the second stage water movement takes place due to capillary flow and vapour diffusion. Temperature gradients play increasingly important roles in affecting direction of moisture movement in this stage (Lemon, 1956).

3.1.2.3 Temperature; Liquid and Vapour Flow

The process of evaporation and heat transfer involves the equilibration of energies of two systems; the atmosphere and the soil (Wiegand and Taylor, 1962). There are two reactions of equilibria; a quick reaction involving a drop in temperature of the soil surface; and a slow reaction involving a decrease in moisture content of the soil profile.

The temperature depression is the greatest at the evaporation zone due to the latent heat of vaporization. Latent heat requirements for soil water may exceed that of pure water (2472 joule g^{-1}) by as much as 840 joule g^{-1} (Nielsen *et al.*, 1972). The evaporation zone ranges from 0-1 cm in depth for moist and conditions to seven cm in depth for dry conditions of high potential evaporation (Richards *et al.*, 1956; Gardner and Hanks, 1966; Fritton *et al.*, 1967).

Both liquid and vapor water flow will occur from warmer to cooler areas in the soil. The rate of flow is greater than that predicted with Fick's Law and the diffusion coefficient for water vapour in air (Philip and deVries, 1957; Cary, 1965). Vapour flow can occur from a cool soil to warmer conditions due to lower vapour concentrations (Fritton *et al.*, 1967). Cary (1965) found that a thermal gradient of 0.5°C/cm at a soil potential of -7 kPa moved as much water as a pressure gradient of 0.2 kPa/cm. At a soil potential of -46 kPa, the same temperature gradient was equivalent to -25 kPa/cm.

The relative amounts of water moved by a thermal gradient as compared to that concurrently moved in the opposite direction by the matric tension gradient may be in total small, but at certain stages may prove significant. Hanks *et al.* (1967) in a light radiated treatment, found that thermally induced flow became greater than upward flow due to matric potential, after 40 days. A total of 7.6 cm of water moved downward, about 10% of that moved upward (0.7 cm) due to the thermal gradient. Although evaporation in soil does not occur under isothermal conditions, models used by Penman (1941) and Gardner and Hillel (1962) which neglect thermal flow, still provide a fairly good approximation (Staple, 1971), probably due to the small amount involved. Rose (1968) found that vapour flow was not important relative to liquid flow in soils with potentials greater than -1500 kPa.

Richards *et al.* (1956) using chloride tracers concluded that water transfer in the vapour phase below a depth of 10 cm was of negligible agricultural significance. Gardner (1959) postulated that vapour diffusion under isothermal conditions will occur primarily at the soil surface resulting in only slightly increased evaporation rates.

3.1.2.4 Properties Affecting Soil Water Evaporation

Evaporation rates are modified by hysteresis, soil texture, profile discontinuities, cracking, and surface residues.

Bresler et al. (1969) showed that evaporation was directly related to the previous wetting rate, either immediately following infiltration or after redistribution for 4 days. Allowing time for redistribution resulted in decreased evaporation. Gardner and Gardner (1969) found that as the frequency of water application increased, more water was lost to evaporation for similar totals applied. Hank *et al*. (1967) concluded that the average rate of drying is almost entirely determined by water movement within the soil and this is influenced only slightly by temperature gradients.

Texture affects evaporation rates by modifying hydraulic conductivity. Finer-textured soils will remain in the first stage of evaporation much longer than coarse textured soils (Lemon, 1956; Wilcox, 1960; Reddy, 1983). Stewart (1984) cited an example where the second stage of evaporation occurred after 12 mm was lost from a sandy soil and after 20 mm from a fine-textured soil. Top layers of coarse-textured material will impede evaporation (Hillel and Talpaz, 1977). Tillage, after wetting, will effectively terminate the first stage and reduce evaporation from greater depths by as much as 50% by creating larger pores (Willis and Bond, 1971). The surface hydraulic conductivities control the evaporation rates of the profile. Tillage to 7.5 cm reduced evaporation only slightly better as compared to a tillage depth of 2.5 cm (Willis and Bond, 1971). Swelling and crusting from rains will change the pore size distribution of the surface from that underneath, resulting in reduced evaporation (Staple, 1971).

Soil cracks extending deep into the profile can cause serious losses of moisture. Adams *et al.* (1969) with simulated cracks, 30 to 60 cm deep and 1 to 7 cm wide found that turbulent air flow within the cracks accounted for most of the water loss. Fifty to sixty percent of the evaporative losses occurred below a depth of 15 cm.

Applications of surface residues have their greatest effect upon reducing evaporation during the first stage (Willis, 1962). During the first stage there is a linear relationship between initial rates of loss and percentage of surface area covered for windy conditions (Willis, 1962). Gravel and straw mulches reduced evaporation from cracks 85 to 90% with no wind and about 60% with windspeeds of 8.9 m/sec (Adams *et al.*, 1969).

3.1.3 TRANSPIRATION

3.1.3.1 Introduction

Transpiration is a function of soil, plant, and meteorological factors (Lemon *et al.*, 1957). Attempts to explain and predict transpiration require consideration of all these factors. Plants form part of a continuum between the soil and the atmosphere in which water moves from regions of higher to lower potential energy (Gardner, 1960).

The amount of water withdrawn from the soil is dependent upon the steepness of the energy gradient and upon the availability of the soil water.

3.1.3.2 Transpiration and Soil Moisture Loss

During conditions of high soil water availability, transpiration rates are dependent upon meteorological conditions (Lemon *et al.*, 1957; Ogata *et al.*, 1960; Taylor and Ashcroft, 1972). As the soil water is depleted beyond a critical threshold, daily transpiration rates become dependent upon soil hydraulic properties (Ogata *et al.*, 1960). The uptake of water decreases as the potential increases and as the moisture content per unit potential decreases (Peters, 1957; Denmead and Shaw, 1962).

Hydraulic conductivity begins to affect transpiration rates at potentials between -200 and -1200 kPa, dependent upon potential evapotranspiration and texture (Gardner, 1960; Denmead and Shaw, 1962; Feyens *et al.*, 1980). At high potentials, hydraulic conductivity and rooting density is sufficient that required flow can be met without very low plant potentials. When the soil potential is low, the plant potential must be much lower to maintain flow with the reduced hydraulic conductivities (Gardner, 1960).

Denmead and Shaw (1962) found that the length of the period of constant transpiration, the steepness of the decrease in the transpiration rate, and the permanent wilting point were dependent upon the potential evapotranspiration. Examples of various soil depletion curves due to relative transpiration rate is illustrated in Figure 3. The relative transpiration rate is the actual evapotranspiration (AET) divided by the potential evapotranspiration rate (PET).

The horizontal curve, A, represents equal availability of soil water from field capacity to almost the permanent wilting point. This curve has been reported for low PET rates by Denmead and Shaw (1962) and Gardner (1960). At a PET of 1.4 mm day^{-1}, the AET for corn grown in a silty clay loam remained approximately equivalent until a soil potential of -1200 kPa was reached (Denmead and Shaw, 1962).

Curve B, based upon Pierce's (1958) proposal (Denmead and Shaw, 1962) was obtained under 'usual' weather conditions of moderate potential evapotranspiration. Denmead and Shaw (1962), found that moderate potential evapotranspiration rates of 3 to 4 mm/day resulted in equivalent AET rates until a soil potential of -200 kPa was reached, resulting in a curve similar to curve B.

Thornthwaite and Mather (1955) proposed a linear relationship, curve C, based upon observations made for a sandy loam soil under very dry conditions of high radiation intensities. Curves C and D agreed well with the curve obtained for a high potential evapotranspiration of 6.4 mm/day in which AET was equivalent to PET until -30 kPa (Denmead and Shaw, 1962).

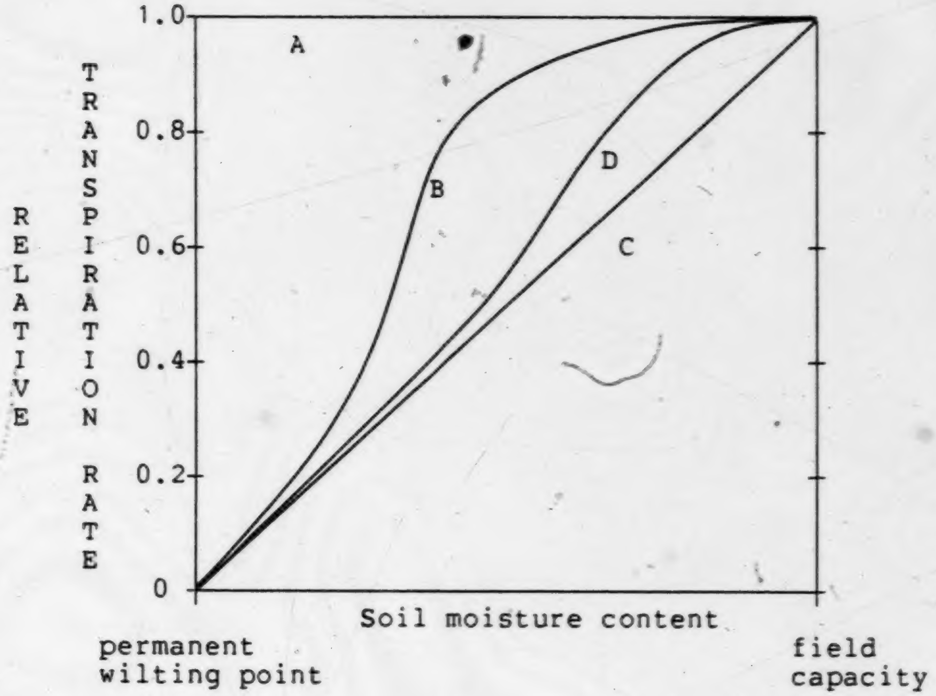


Figure 3. Relationships between relative transpiration rate and soil moisture

(adapted from Denmead and Shaw, 1962)

The permanent wilting point is a function of the magnitude of the potential evapotranspiration and texture. Gardner (1960) found that the wilting point varied from 16 to 23% for a clay and from 23 to 34% for a silty clay loam soil for PET rates of 1.4 to 6 mm day^{-1} (Denmead and Shaw, 1962).

3.1.3.3 Rooting Distribution and Soil Water Depletion

As soil water becomes limiting, transpiration rates become dependent upon the rooting distribution and hydraulic conductivity of the soils (Gardner, 1960; Ritchie *et al.*, 1972). Gardner (1960) found that the rate of uptake from a given soil is proportional to the effective length of roots.

Given initial conditions of a uniformly moist profile, roots will gradually elongate and proliferate down the profile as the shallower layers dry (Ogata *et al.*, 1960; Draycott and Durrant, 1971; Belmans *et al.*, 1979). Water is removed more quickly from the surface layers due to denser rooting and evaporation (Ogata *et al.*, 1960; Belmans *et al.*, 1979). Generally there is a gradual decrease in rooting density with depth (Gardner, 1960; Belmans *et al.*, 1979); however, frequent but light irrigations (Rose and Stern, 1967) or a shallow water table (Belmans *et al.*, 1979) can result in a shallower, more dense network of roots.

3.1.3.4 Physiological Factors

The stage of plant maturity will affect transpiration rates (Denmead and Shaw, 1962; Bowman and King, 1965; Ritchie *et al.*, 1972; Ritchie, 1980a). Leaves of corn and most small grain crops become senescent at a certain stage of maturity regardless of the soil water status, resulting in a decline in transpiration (Bowman and King, 1965; Ritchie *et al.*, 1972). Stress-induced maturity, due to nutrient (Brown, 1971; Ritchie, 1980a) or water deficiencies (Ritchie *et al.*, 1972; Garrity *et al.*, 1982) can become confused with limited soil water that also causes reduction in transpiration. Water stress, regardless of time of occurrence during crop growth, can reduce water use efficiency throughout the growing season (Garrity *et al.*, 1982).

Root impedance to transport can be an important factor determining extraction patterns from the soil (Wind, 1955; Ogata *et al.*, 1960). Plant resistance to transport is much greater than soil resistance throughout a wide range in soil water content (Taylor and Klepper, 1976). The resistance will vary due to moisture content, age of root, rooting density, and physiological stage of plant (Hillel *et al.*, 1976)

3.1.4 DRAINAGE

3.1.4.1 Field Capacity

Soil water held between saturation and field capacity is subject to gravity induced flow. In the absence of saturated conditions or a shallow water table, this is referred to as redistribution, otherwise, it is internal drainage (Hillel, 1980b). Water that has 'drained' below the root zone is the drainage component in the soil water balance.

Field capacity is that soil moisture content at which excess gravitational water has drained and downward movement of water has ceased (Veihmeyer and Hendrickson, 1931). The importance of field capacity is the concept of a specific water retention value opposing the force of gravity (Peters, 1965). The exact point at which field capacity occurs at, however, remains difficult to define as drainage has been shown to go on indefinitely at ever decreasing rates (Nielsen *et al*., 1973) down to potentials as low as -50 kPa (Hillel, 1980b).

Common laboratory measurements of field capacity are conducted on 2 mm sieved, oven dried, samples at 33 kPa (Peters, 1965). Cavazza *et al*. (1973) found that -30 kPa on sieved samples gave values similar to average field obtained values. Field measurements, however, indicate that field capacity occurs at potentials between -4 and -10 kPa (Russell, 1961; Webster and Beckett, 1972; Parkes and O'Callaghan, 1980) for most textural ranges. Webster and Beckett (1972) found that the moisture status of freely drained sandy loam soils stabilized after 2 to 3 days at -3 to -7 kPa, that for loams and clay loams after a week at -4 kPa, and that clays had no detectable transition point of stabilization. Values of -5 kPa were suggested for clays as flow was almost undetectable at this value. Laboratory measurements on undisturbed soil cores at potentials of -10 kPa (Cavarazza *et al.*, 1971) and -5 kPa (Hall et al. 1977) have been found to agree with field measurements. Hillel (1980a) reports drainage to occur to about -50 kPa.

3.1.4.2 Properties Affecting Drainage

The main physical property affecting redistribution rates and amounts is the pore size distribution as influenced by texture and bulk density. As particle size increases there is an increase in pore size but a decrease in total porosty (Hall *et al.*, 1977). For sandy soils drainage is initially rapid but suddenly decreases due to the narrow distribution of pore sizes (Hillel, 1980b). Redistribution rates in fine- and medium-textured soils continue at gradually declining rates for days and sometimes weeks due to the wider distribution of pore sizes.

The drainage process consists of two parts; an inital rapid throughflow of water to deeper portions of the profile, followed by a "normal" drainage process. This may be due to preferential routes caused by vegetation (Kanchanasut and Scotter, 1982), macrofauna, and structural cracks (Quisenberry and Phillips, 1976; Parkes and O'Callaghan, 1980). Quisenberry and Phillips (1976) found that 40% of applied water with a chloride tracer penetrated below 90 cm within 1 hour following irrigation of silty loam and silty clay loam soils.

Actively growing vegetation can affect redistribution rates. Wilcox (1960) found that evapotranspiration was occurring largely at the expense of free water that would have normally drained away. Vegetation through interception and stemflow will reduce the amount of water reaching the soil (Kanchanasut and Scotter, 1982).

Any profile discontinuity that results in a change in pore size distribution will result in lower drainage rates. For water to flow into an underlying coarser layer, the moisture content in the overlying layer must increase until the matric tension reaches the air-entry value of the larger pores (Miller, 1973; Gardner, 1979). At this point the profile will seem to suddenly 'empty' because of the increased moisture flow into the coarser layer (Hillel 1980b). The presence of finer layers on the other hand, will impede flow due to smaller pore sizes.

3.1.5 EFFECTS OF A SHALLOW WATER TABLE

3.1.5.1 Introduction

The presence of a shallow groundwater surface affects the soil water balance of the root zone through contribution of water to transpiration and evaporation. The upper boundary of the capillary fringe from the water table is defined as the air-entry value of the soil and the lower boundary, the water table, is where the pressure of the groundwater equals atmospheric pressure (Hillel, 1980b).

3.1.5.2 Drainage

In the presence of a shallow water table, the addition of small amounts of water can result in a very large rises in the water table (Freeze and Cherry, 1979; Gillam, 1984). Meyboom (1967) found that a heavy rainstorm caused a rise in the water table by as much as 20 times the depth of rainfall. The larger than expected rise was interpreted as due to air entrapment (Freeze and Cherry, 1979). When the capillary fringe extends to the ground surface, the addition of a very small amount of water can result in an immediate and large rise in the water table. Gillam (1984) reported a 30 cm rise in a water table in 0.25 min from the addition of 0.3 cm of water.

3.1.5.3 Evaporation

Gardner (1958), in a theoretical study of steadystate upward flow from a water-table to an evaporation zone, and Gardner and Fireman (1958), in a laboratory study, showed that the evaporation rate can be limited either by the potential evaporation or by the maximal rate at which the soil can transmit water, whichever is less. Where the water table is near the surface, the potential at the soil surface is low and the evaporation rate is determined by external conditions. With increasing depth of the water table, the matric potential at the soil surface decreases and upward flow becomes more limited by soil properties (Hadas and Hillel, 1968).

The maximum water table depth at which evaporation is still externally controlled and not limited by soil properties is referred to as the 'critical water table depth' (Anat *et al.*, 1965). At this point the pores begin to rapidly desaturate with a decrease in soil potential. Hillel (1980b) referred to this value as the air-entry value corresponding to the top of the capillary fringe. Anat *et al.* (1965) cited the critical water table depth for fine sand as 60 cm. The finer the texture of the soil, the higher the capillary fringe and the greater the evaporation rate for given depths (Gardner, 1958). The actual amounts evaporated, however, are dependent upon the depth of the water table and the potential evaporation (Wind, 1955; Gardner and Fireman, 1958; Gardner, 1958).

Soil horizonation affects evaporation rates. The amount of the effect is related to both the type of profile discontinuity and the depth of the water table (Willis, 1960). The existence of a fine layer over a coarse one has a relatively small effect upon the evaporation rate, regardless of the water table depth (Willis, 1960). The evaporation rates in this case will approximate that of a profile entirely composed of the fine soil. A coarse layer over a fine layer, however, will result in very large differences depending upon the thickness of the coarse layer. The thicker the layer, the greater the evaporation rate for a specific water table depth. As depth to the water table increases, the presence of layers has less effect on the evaporation rates.

Several researchers have studied the effect of PET upon the actual evaporation rates in the presence of a water table (Schlausener and Corey, 1959; Anat *et al*., 1965; Hadas and Hillel, 1968). Initially increasing PET resulted in linear increases in AET. When PET reached a certain value, AET decreased with further increases in PET. The deeper the water table and the coarser the soil, the smaller the potential evaporation value at which the decrease occurred(Hadas and Hillel, 1968). Schlausener and Corey (1959) concluded that high PET removed water from the surface layer faster than the conductivity rates could replace it from deeper layers. Anat et al. (1965) found that whenever this occurred hysteresis and the evaporation rates were reduced by 20 to 50%. Hadas and Hillel (1968), by measuring evaporation under diurnal conditions of high evaporativity, found lower than expected evaporation rates, attributing the difference to hysteresis.

Under certain conditions the evaporation rate is lowered due to the creation of a two-layer condition in which a dry surface layer acts as a "diffusion barrier" to vapour movement (Hadas and Hillel, 1968).

Hellwig (1978) reported the presence of two daily peaks in evaporation rates from a sand lysimeter with a water table near the surface. One peak occurred at sunrise and was related to air temperature and the presence of a condensation surface. It was independent of the water table depth until the water table was below 40 cm. The other peak occurred later in the afternoon and appeared to be related to radiation. The deeper the water table the less the absolute value of the peak and the greater the lag.

3.1.5.4 Transpiration

Gardner (1958) theorized that in the presence of an actively transpiring crop the potential at the bottom of the root zone may be taken as the upper boundary for a shallow water table. Water uptake for a soybean crop in the presence of a shallow water table was found to be related to a small number of roots near the capillary fringe (Reicosky *et al.*, 1972). Taylor and Klepper (1975) found that ryegrass roots were denser and shallower in soils with a shallow water table.

Plant roots create drier conditions lower in the profile and result in an upward flux of water. Upward fluxes of 4 mm/day and 2 mm/day have been reported by Van Bavel *et al.* (1968b) and Stone *et al.* (1973) in the absence of a water table. Purvis (1964) studied winter wheat and sugar beet production on sandy soils with shallow water tables in New Jersey. With depth to the water table varying between 100 to 300 cm, 40 to 70% of the crop water requirement was supplied by groundwater. Saini and Ghildyal (1978) in Northern India found that 36 to 73% of the total water requirement for winter wheat was met by upward flux from groundwater through a silty clay loam. A study by Read and

Pohjakas (1981) in southern Alberta with a barley crop found that groundwater contributed 21.5% to AET for a loamy sand and 30.5% for a clay loam. The water table was at 180 cm. Fallow soils under similar treatments did not have any groundwater contributions to evaporation.

3.1.6 SOIL WATER BALANCE EQUATION

A simplified form of the soil water balance equation is:

$$P = Et + D + \Delta S \tag{3}$$

where P = precipitation,
Et = evapotranspiration,
D = drainage, and
ΔS = change in soil moisture.

This equation may be rearranged to solve any individual component or combination of components. Usually the more difficult to measure components, D and Et are determined by difference:

$$Et + D = P - \Delta S \tag{4}$$

after measuring P and ΔS.

Equation (4)is commonly used by many researchers where no runoff or interflow occurs. For soil profiles where there is no drainage, the equation may be reduced even further (Shouse *et al.*, 1982; Stewart, 1984):

$$Et = P - \Delta S \tag{5}$$

Shouse (1980) assumed that D was equal to zero, as the observed hydraulic gradients were small and the water content below the root zone was low and did not change significantly during the season.

If drainage occurs, Et and D must be separated. Holmes (1956, 1964) solved the equation for Et, by measuring ΔS with a neutron probe, D by outflow from a lysimeter, and obtained P from meteorological methods. Rose and Stern (1965) determined Et by using a neutron probe and measuring the hydraulic conductivity to estimate the D component. They later used the hydraulic conductivity function (Rose and Stern, 1967) to separate evaporation (E) from transpiration (T). Errors originating from using a laboratory moisture characteristic curve to infer field potential and from spatial variability of hydraulic conductivity were postulated as contributing inaccuracies of up to 47%. If the D component was neglected, they concluded that an error of up to 16% in T would occur.

The presence of a water table shallow enough to contribute moisture to the root zone or to the evaporating surface can make the water balance difficult to solve for Et. Nikolski (1977) and Saini and Ghildyal (1978) solved for the irrigation requirement (I) for crops in the presence of a shallow water table inferring upward flow from (U) the hydraulic conductivity:

$$I = Et = P - D + U + \Delta S \qquad (6)$$

McGowan **et al.** (1980) estimated *Et* for a watershed by measuring *P*, runoff from streamflow and ΔS from neutron measurements. They found that the main sources of error in measuring *Et* during wet years were the neutron probe and the estimation of drainage and during dry years from the variability of rainfall. Combining these errors the standard error of annual *Et* (372 to 440 mm) was ±30 mm, regardless of the amount of rainfall.

Soil moisture and changes in soil moisture (ΔS) may be determined indirectly from estimates of evapotranspiration. Thornthwaite and Mather (1955) estimated periods of moisture deficiency and excess by comparing calculated *PET* with measured *P*. Baier and Robertson (1966) used estimates of the water holding capacity and *PET* along with plant growth coefficients to obtain estimates of soil moisture from meteorological records. Spittlehouse and Black (1981) used a similar approach to estimate forest soil moisture and losses in soil moisture due to either drainage or evapotranspiration.

The soil water balance equation can be solved for ΔS by using meteorological data and empirical relationships between the components (Reddy, 1983). For practical purposes; however, it remains simpler to solve the equation for *Et* by measuring *P* and ΔS. Drainage may be measured or inferred by a variety of methods explained in the following section.

3.1.7 MEASURING TECHNIQUES

3.1.7.1 The Neutron Probe

The change in soil moisture content (ΔS) is perhaps the most important and simplest of the soil water balance components to obtain. The accuracy of the determination of the change in soil moisture with time is related to the magnitude of the change and to instrument resolution. The magnitude of change in soil moisture content (ΔS) is dependent on the duration of time between samples, soil retention properties, and the magnitude of the other balance components. Instrument resolution is an inherent property of the instrument itself.

A commonly used instrument for soil water studies since the late 1950's is the neutron probe. This instrument has proven to be a reliable and accurate method for obtaining sequential readings at the same location and determining a water balance.

The basic principle behind the neutron probe is the emittance of fast neutrons and the detection of neutrons that have been slowed down through collision with H molecules. The number of slow neutrons measured (the count) for a specific time period is an indication of the moisture content. The distribution of counts about the mean for a standard medium is normal (Milanova, 1969). Most neutron probes reported in the literature have a standard deviation of less than 1% for 1 to 5 replicate observations (Holmes and Colville, 1964; Milanova, 1969; Sinclair and Williams, 1979; and McGowan and Williams, 1980a). Standard deviations of about 2% have also been reported (Nixon and Lawless, 1960; and De Boodt *et al.*, 1969). There are indications that a change in technology has improved the resolution ability of the neutron probe as the older probes did not possess an annular source surrounding the counter-tube, thus contributing to variation due to geometrical placement (Holmes, 1984).

The accuracy gained by increasing the count time is very little. Bowman and King (1965) found that count times past two minutes did not significantly add to precision. Sinclair and Williams (1979) and McGowan and Williams (1980a) used 16 s count periods. Sinclair and Williams (1979) did not find any significant increase in precision by increasing count time past 20 s.

In a field experiment by Holmes and Colville (1964, 1970) it was found that for a profile water content of 50 cm, the neutron probe had a standard deviation of 0.08 cm and that for a profile 2.25 m deep a water gain or loss of 0.5 cm could be resolved. Bowman and King (1965) reported weekly errors of 3.8 mm (15 to 21%) for the determination of *Et* and a 3 month error of 15.7 mm (about 6% assuming a daily *Et* of 3 mm). Van Bavel *et al.* (1968b) concluded that with *Et* rates of 5 to 9 mm/day a neutron probe could resolve changes in moisture for measurements at least 5 to 7 days apart. If determinations of soil water change for periods less than one week are desired, the weighing lysimeter is recommended (van Bavel *et al*., 1968b; Holmes, 1984).

3.1.7.2 Separation of Evapotranspiration and Drainage

Introduction

In soil water balance studies, separation of the moisture lost due to drainage and evapotranspiration is crucial to the correct evaluation of ΔS. As there is no direct method of measuring actual evapotranspiration, many methods rely upon the determination of the drainage component.

The following methods have been used to determine drainage rates and thus to separate drainage from evapotranspiration:

(i) the field capacity concept,
(ii) flux/instantaneous profile method,
(iii) zero flux plane,
(iv) gradient method,
(v) tracers such as chloride and tritiated water,
(vi) empirical descriptions, and
(vii) lysimeters.

Selection of the appropriate method will depend upon the relative limitations and advantages of the specific method in accordance with the objectives, required accuracy, and economics of the experimental design.

Field Capacity Concept

At field capacity, the amount of soil water lost from the root zone due to drainage is insignificant in terms of total soil moisture storage. The use of field capacity for demarcating the moisture content at which drainage ceases is perhaps the simplest approach, but also the most error prone due to the fact that drainage never actually ceases. Drainage rates at -5 kPa potential (Webster and Beckett, 1972), 1/10 potential evapotranspiration (Hillel, 1980b) and 0.1 cm/day (Ritchie, 1980a) have been suggested as practical field measurement values at which point drainage can be neglected.

Errors of 20 to 30% in calculation of evapotranspiration can occur if drainage is not considered (Robins *et al.*, 1954; van Bavel *et al.*, 1968a; Rouse, 1969).

Flux/Instantaneous Profile Method

This method involves the use of the continuity equation with measured unsaturated hydraulic conductivities and a characteristic curve to calculate actual flow. Determination of the hydraulic conductivity-moisture content relationship may be conducted on undisturbed soil columns in the laboratory (Gardner, 1956; Klute, 1965), from the Millington-Quirk theory (Millington and Quirk, 1959; Nielsen *et al.*, 1973) or from actual *in situ* field measurements using the instantaneous profile method as described by Richards and Weeks (1953) and Hillel et al. (1972).

Although laboratory methods offer the advantage of controlled conditions and are relatively economical, they do not reflect field conditions.

The instantaneous profile method represents field conditions and gives reasonably accurate results (Richards *et al.*, 1956; Rose and Stern, 1965, 1967; van Bavel *et al.* 1968ab; Nielsen *et al.*, 1973). This method involves flooding a plot to achieve unit hydraulic gradient, covering it to prevent evaporation, and then measuring water content and soil potential frequently and simultaneously. Hydraulic conductivity may be calculated for the moisture content or the soil potentials encountered during drainage, 0 to -50 kPa (Hillel, 1980a).

Advantages of this method are (Klute, 1973; Hillel 1980a): it does not assume uniformity of the hydraulic properties; boundary conditions need not be established; hysteresis can be accounted for, although with difficulty; time and effort are less than that required in the laboratory; and once the hydraulic conductivity function is established for a site, it can be extended to vegetated conditions.

Several limitations of the instantaneous profile method have been noted. Some studies (Wilcox, 1960; van Bavel *et al.*, 1968b) noted that it did not give adequate results for estimation of evapotranspiration; it loses reliability with more frequent and copious water applications (van Bavel *et al.*, 1968b); diurnal temperature fluctuations can affect water movement in the top 15 to 25 cm (Klute, 1973) and hydraulic gradients within the root zone (Van Bavel, 1968b); it relies upon the assumptions of no lateral flow and no lateral variation in hydraulic conductivity, moisture content, and soil tension (McGowan and Williams, 1980a); errors may occur when rapid drainage occurs in structural cracks and worm burrows, especially when evapotranspiration is taking place (Parkes and O'Callaghan, 1980); and the anormal distribution of unsaturated hydraulic conductivity values can lead to inaccurate representation by the mean (Hartman *et al.*, 1980).

Zero-flux Plane

Where evapotranspiration takes place from the root zone and where drainage takes place from the bottom of the root zone, a plane exists between the two where the hydraulic gradient is zero. This plane is referred to as the zero-flux plane or static zone (Giesel et al., 1970). Above this plane water movement is upward and below it is downward. During vegetative growth the zero-flux plane is often found below the root zone (Kreutzer *et al.*, 1980).

Several studies have compared the zero flux plane method to others. Hartmann *et al.* (1980) compared it to field hydraulic conductivity data and found that the conductivity method gave either much higher or much lower seepage values, with the mean value being too high. They concluded that the zero flux plane method for determination of water balance was more representative of field conditions given the typical log-normal distributions of hydraulic conductivities. McGowan and Williams (1980a) compared the zero flux plane method to the gradient method and found good agreement. Drying depths inferred from hydraulic potentials, however, tended to be slightly deeper than those inferred from the neutron probe. This was attributed to differences in sensitivity between the two techniques.

Gradient Method

This method has been developed and used by McGowan (1974) and McGowan and Williams (1980a). It involves examining graphs of water content versus time at specific depths for irregularities attributable to root extraction, rainfall, and harvesting. Graphs of water content versus time often show initial slow, almost non-existent rates of water loss equated to loss by drainage, followed by a sudden discontinuity interpreted as root extraction. Ogata and Richards (1957), Wilcox (1960), and Black *et al*. (1969) have indicated that drainage rates will decrease consistently with time (barring profile textural changes). The discontinuity will increase in depth during the growing season as the roots grow. Water losses before the discontinuity occurs are attributable to seepage, whereas those after are inferred to be due to evapotranspiration.

This method is based upon two assumptions: (1) that there is no loss of water by downward drainage after the discontinuity, and (2) that roots do not extract a significant amount before the discontinuity (McGowan and Williams, 1980a). These assumptions were checked during a field study by McGowan and Williams (1980a) where they obtained a good linear fit of the gradient method with the zero-flux plane method. McGowan (1974) found for spring cereal crops with expanding root systems that the discontinuity depths were about 10 cm deeper than actual root depths. He indicated that these differences in depths would not introduce any significant errors into water balance calculations.

Tracers

Dyes, chlorides, bromides and radioactive substances, such as tritiated water, are commonly used to 'trace' water movement. The general method is to add a known amount of the labelling material with a known amount of water and then after specified periods of time simultaneously measure water content and 'tracer' concentration at various depths.

Richards *et al.* (1956) obtained a 7% difference in the estimation of evapotranspiration between the hydraulic gradient-conductivity method and that using chloride. They noted two factors which may influence the chloride concentration in soil-water; the removal of chlorides through distillation, condensation, and upward film flow; and negative adsorption as the moisture content of the soil is depleted. Quisenberry and Phillips (1976) successfully used chlorides to follow deep and rapid seepage associated with structural cracks. For slower drainage rates that occurred after the structural seepage, water and chloride concentrations did not correspond for the top 15 cm, but did for the rest of the profile. They attributed this to profile discontinuities associated with the tillage layer. Kanchanasut and Scotter (1982) used bromide to study drainage in vegetated plots. They suggest that the vegetation, by inducing preferential flow pathways, retarded the leaching of bromide from the soil near the surface.

Tritiated water has been used in several studies (Woods and O'Neal, 1965; Kreutzer *et al.*, 1980). Woods and O'Neal (1965) injected tritiated water at several depths in the root zones of small trees to measure transpired water. Rose and Stern (1967); however, suggest that this water movement within the profile may lead to errors in interpretation. Kreutzer *et al.* (1980) injected tritiated water at 60 cm depth and compared water balance estimations by this method with that of the zero flux plane. Inaccurate results were obtained if the tracer distribution was within the zone of active roots.

Empirical Descriptions

Numerous researchers (eg. Richards et al., 1956; Black et al., 1969; Aston and Dunin, 1977) have found that after infiltration ceases, the moisture content is inversely proportional to time (Equation 7).

$$\theta = aT^{d} \qquad (7)$$

where θ = water content
T = time

a,d = constants, where d is related to diffusivity (Hillel, 1980b).

The relationship is linear on a logarithmic scale. linear on a logarithmic scale. Use has been made of this equation to calculate vertical flow velocity and estimate hydraulic conductivity. Reasonably good agreement between laboratory and field values were obtained. These models, however, seem to be best suited for homogeneous soils. Fitting such drainage functions to layered soils might prove difficult.

Lysimeters

A weighing lysimeter with leachate collector is the most accurate of methods (± 0.2 mm per day) but it lacks representation and flexibility of field variability (Holmes, 1984). When combined with neutron access tubes moisture distribution may also be described.

3.2 MATERIALS AND METHODS

3.2.1 SITE CHARACTERIZATION AND PREPARATION

Soil at the study site was characterized using a soil pit located approximately 80 m to the south at the top of the catena previously described. The pit is part of the same local rise on which the study site is located. A profile description of this pit is provided in Appendix A.

The site was tilled, during May 8 to 10, 1983, and the access tubes installed May 11 and May 17 to 19. The plots were tilled again to remove tracks left by the coring truck and the barley plots were seeded on May 26.

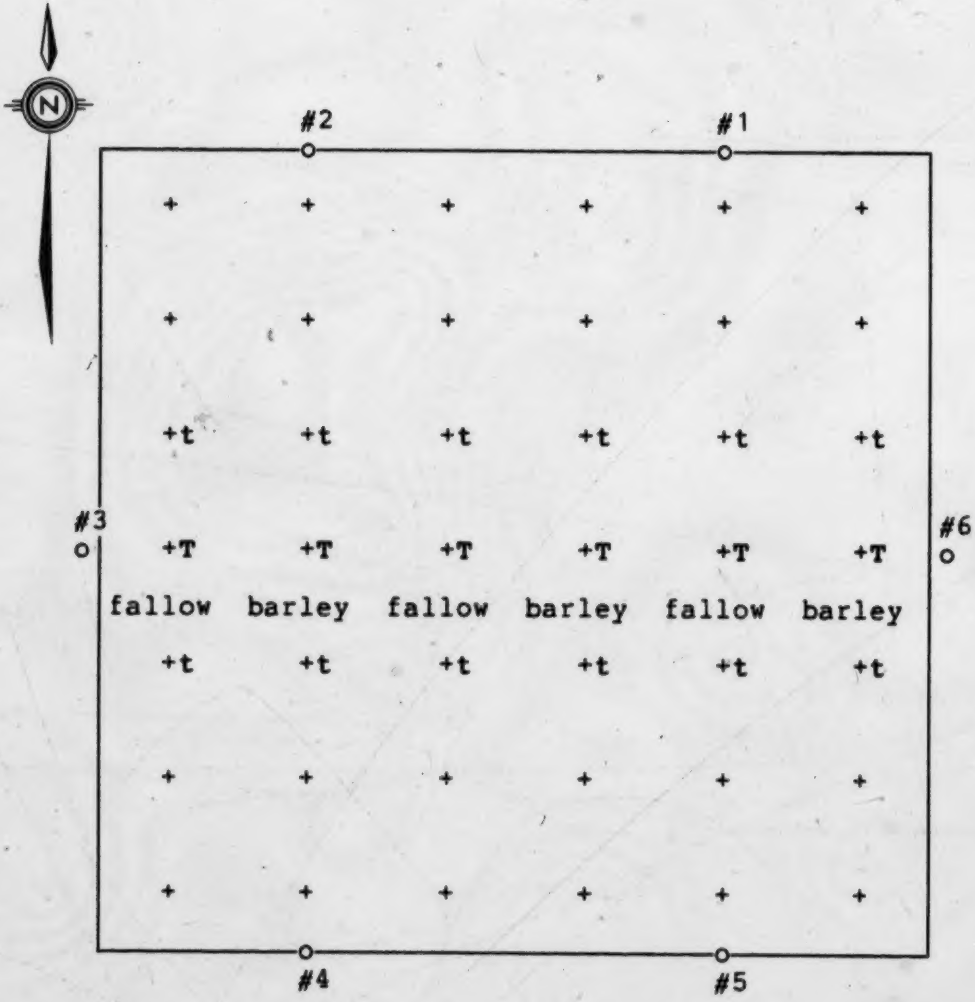
Roundup herbicide was sprayed twice on the fallow plots during the summer, during June and August, to control Canada thistle. The fallow plots were tilled twice, in July and August.

Surface elevations at the neutron access tube locations and the water well locations were determined.

3.2.2 SAMPLING AND ACCESS TUBE INSTALLATION

Aluminum access tubes for the neutron probe were installed at each grid location (Figure 4) with a truck-mounted hydraulic coring unit. The coring tube extracted a soil core 4.5 cm in diameter and 100 cm long. The aluminum access tubes fit snugly into the cored holes. Optimum moisture conditions at the time of coring resulted in very little observable compaction of the core. The core surface was usually depressed only about 1 cm and never more than 3 cm.

The cores were examined to determine the depth of the Ah horizon and for presence of sand lenses. They were then sectioned into 10 cm increments from 0 to 40 cm and 20 cm increments from 40 to 100+ cm and bagged for subsequent laboratory analysis.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barley
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barley
fallow
barley
#4
#5

+ access tube location
o water well site
T main tensiometer nest
t secondary tensiometer nest

Figure 4. Instrument locations in study plots

3.2.3 LABORATORY ANALYSIS

Bulk density, 5 point particle size analysis and the moisture characteristeric curve were conducted on soil samples for depths listed in Table 2. The soil was weighed while moist, dried at 105°C, and then weighed dry to determine gravimetric moisture content and bulk density. The soil was then ground and sieved in a rotary drum with 2 mm openings. The Ah samples were hand ground and sieved.

Particle size analysis was performed using the hydrometer method (McKeague, 1978) except for the clay fraction which was separated using the pipette method (Jackson, 1979). All Ah samples were pretreated with hydrogen peroxide to remove colloidal organic matter (McKeague, 1978). Samples below 60 cm were pretreated with the addition of 5 ml of 1N HCl to remove carbonates.

Soil moisture characteristic curves were determined using a pressure plate apparatus (Richards, 1965) on disturbed samples. Mass water content was determined at pressures of 33, 100, 300, 1000, and 1500 kPa.

3.2.4 FIELD INSTRUMENTATION

A Campbell Pacific Nuclear Neutron Probe (Model 503) was used to determine moisture content at each of the 42 locations on the plot. The first moisture measurements were made on May 26, 1983 and were measured weekly until September 8. Measurements were made at 15 cm increments from 15 cm to 90 cm. Total profile water was calculated to a

TABLE 2. Depth intervals of soil for analysis

BULK DENSITY (cm)	PARTICLE SIZE (cm)	PRESSURE PLATE (cm)
0-10		
10-20	0-20	0-20
20-30		
30-40	20-40	20-40
40-60	40-60	40-60
60-80	60-80	60-80
80-100+	80-100+	80-100+

depth of 1 m.

Mercury manometer tensiometers were constructed using porous cups, (6 cm long, 1.9 cm O.D., and 0.24 cm wall thickness) with a bubbling pressure of about 200 kPa (Soilmoisture, 1974), glued onto polyvinylchloride tubing. Nylon manometer tubing connecting the water source to the mercury had dimensions of 0.24 cm O.D. with a 0.04 cm wall. Tensiometer nests were established at three locations within each plot (Figure 4). A main nest was located adjacent to the access tube at the center of each plot and secondary nests were situated adjacent to the access tubes on either side of the main nest. The main nests consisted of tensiometers at depths of 30, 45, 60, 75, and 90 cm, whereas the secondary nests had tensiometers only at 30 and 60 cm. Readings were taken concurrently with the neutron moisture readings and were taken at 8 a.m. to avoid temperature effects.

Six water wells to depths of 2 m were established at various perimeter points at the site (Figure 4). Readings were taken during the same day as the other readings.

Precipitation, temperature, relative humidity, wind, pan evaporation, and solar radiation data were obtained from a meteorological station situated about 20 m to the east of the site. Net radiation and total radiation were measured with a CSIRO pyradiometer and a KIPP CM6 pyranometer at the Edmonton-Stony Plain station respectively. The Edmonton-Stony Plain station is located about 50 km to the west of the study site at lat. 53° 33' N, long. 114° 06' W.

3.2.5 POTENTIAL EVAPOTRANSPIRATION DETERMINATION

Three methods were used to calculate the potential evapotranspiration (*PET*); the Class A Evaporation Pan located at the Ellerslie meteorological station; Penman's equation; and a multi-regression technique by Baier and Robertson (1965).

3.2.5.1 Pan A Evaporimeter

The Pan A evaporimeter is a standard device for measurement of free water evaporation by Canadian and U.S. weather offices. It is constructed of galvanized steel and is 1.21 m in diameter and 0.25 m deep (Gray *et al.*, 1970). Sonmor (1963) suggested a coefficient of 0.64 cm cm^{-1} for estimating the consumptive use of barley for maximum yield in Southern Alberta from Class A Pan data (Gray *et al.*, 1970).

3.2.5.2 Penman Method

The general form of the equation is:

$$LE = \frac{(\Delta/\gamma)J + LEa}{(\Delta/\gamma) + 1} \qquad (8)$$

where LE is potential evaporation, J is the heat budget, LEa is an expression for the "drying power" of the atmosphere, Δ is the slope of the saturation vapor pressure versus the temperature curve at mean air temperature and γ is the psychrometric constant (Heapy, 1971; Hillel, 1980b).

Delta(Δ)

The Goff-Gratch formula as given by List (1958) was used for calculating Δ.

Psychrometric Constant(γ)

γ is the constant (0.65) of the wet and dry bulb psychrometer equation. Calculations based on data presented by List (1968) revealed that this assumed value of γ is without error at a wet-bulb temperature of 45°F and the actual error would seldom exceed 1 per cent (Heapy, 1971).

Heat Budget(J)

The heat budget can be divided into its component parts by the relationship (Penman, 1963):

$$J = J_s(1-r) - J_1 \tag{9}$$

where J_s is incoming shortwave radiation, r is surface albedo, and J_1 is net outward long-wave radiation.

Incoming short-wave radiation (J_s) was measured with an Eppley pyrheliometer. The radiation is reported in MJ m^{-2} and must be converted to cal cm^{-2} by multiplying by 23.883

to be used in formula (9).

Outgoing long-wave radiation (J_1) was estimated from a formula presented by Penman (1963):

$$J_1 = \sigma T^4(0.56-0.09e_d^{1/2})(0.10+0.90\ n/N) \qquad (10)$$

where σ is the Stefan-Boltzmann constant, (1.98×10^{-9} mm $H_2O/cm^2/day/°K^4$); T is the mean air temperature (°K); e_d is the actual vapour pressure of the air (mm Hg), and n/N is the ratio of actual to possible hours of bright sunshine. The actual vapour pressure was calculated from Equation 13. Possible hours of bright sunshine was taken from List (1958) for Latitude 54°N.

Surface albedo (r) is the portion of incident solar radiation that is reflected from the soil and crop. Verma (1968) used a r value of 0.25 for a green crop in the Edmonton area. Heapy (1971) used 0.18 for barley and 0.12 for bare soil in a study at Ellerslie. The bare soil albedo was used for the fallow plots and in the barley plots until full emergence, then 0.18 was used for the albedo for the barley plots.

Drying-power of the atmosphere(*LEa*)

The equation for the drying-power of the atmosphere, appropriate to a crop surface, is given as (Penman, 1963):

$$LEa = 0.35(1 + u/100)(e_s - e_d) \qquad (11)$$

where u is windspeed (miles/day) at a height of 2.0 m, e_s is saturation vapour pressure and e_d is actual vapour pressure (mm Hg). Wind data from Ellerslie at 10 m was converted to a wind speed at 2 m using the following power law:

$$u_1/u_2 = (z_1/z_2)^m \tag{12}$$

where u_1 is the wind speed at height z_1, u_2 is the wind speed at height z_2 and m is a variable depending upon the stability of the air layer. A value of 0.2 was used for m (Heapy, 1971).

Saturation vapour pressure (e_s) was computed by the equations of Goff and Gratch presented by List (1958) in the Smithsonian Meteorological Tables. A linear correction was applied for actual air pressure as moist air does not exactly fulfill relationships that express the ideal gas law (Harrison, 1965a; cited by Heapy, 1971). The Edmonton average pressure of 935 mb was used for a correction factor (Heapy, 1971).

The actual vapour pressure (e_d) was calculated from e_s and the relative humidity (RH) using the following formula:

$$e_d = e_s \times RH/100 \tag{13}$$

3.2.5.3 Baier-Robertson Estimation of Potential Evaporation

This technique for estimating daily potential evaporation requires only simple meteorological observations and astronomical data. The technique was developed by Baier and Robertson (1965) using data collected from a five year period, 1953 to 1957, during the months May through September from six meteorological stations across Canada. Nine variables were studied and simple and multiple linear correlation and regression formulae were developed. Daily records of latent evaporation as the dependent variable was correlated with daily data of several meteorological and astronomical parameters as independent variables. These include:

1. **Latent Evaporation**(LE). This was measured with a Bellani plate atmometer. The formula given by Baier and Robertson (1965) solves LE in cm^3/day from a Bellani plate atmometer and must be multiplied by 0.08636 to be converted to mm/day.
2. **Maximum temperature**(T_m). The temperature used in the formula is in degrees F.
3. **Temperature range**(T_d). The difference (°F) between daily maximum and minimum temperatures.
4. **Wind**(u). Total daily wind in miles at a height of at least 1.5 m.
5. **Duration of bright sunshine**(n). Daily sunshine in hours.
6. **Vapour pressure deficit**($e_s - e_d$). e_s was determined from formulae presented by List (1958) as described in the previous section on Penman's method.

7. **Solar energy at the top of the atmosphere.** (Q_a). Total daily solar radiation in cal/cm² from meteorological tables in List (1958).
8. **Daylength** (N). Possible hours of sunshine from meteorological tables in List (1958).
9. **Total sky and solar energy on a horizontal surface** (Q_t). Total daily sky and solar energy in cal/cm² was calculated using the following relationship:

$$Q_t = Q_a(0.251+0.616(n/N)) \qquad (14)$$

(Baier and Robertson, 1965)

Baier and Robertson (1965) offered eight possible multiple correlation methods each using various combinations of the variables. The method used in this study utilized the most number of variables and was judged by Baier and Robertson (1965) as yielding the most accurate results. The multiple correlation formula used was:

$$\begin{aligned} 0.08636LE = {} & -53.39 + 0.337T_m + 0.531T_l + 0.00107Q_a \\ & + 0.00512Q_t + 0.00977u + 1.77(e_a - e_d) \end{aligned} \qquad (15)$$

Daily LE was calculated for the months May through September.

3.2.6 DRAINAGE ESTIMATION

Of the methods described in the literature review, the following methods were utilized; field capacity, zero flux plane and the gradient method. Field capacity values determined by tensiometers at -5 kPa and -10 kPa were used.

The assumption was made that potential evaporation as determined by the Penman method, corrected for evaporation from a free water surface using a coefficient of 0.8, was equivalent to water lost from the soil when the soil potential was above -5 to -10 kPa. Also moisture differences between barley and fallow plots were tested for significant differences at specific depths to determine when root extraction began.

3.3 RESULTS AND DISCUSSION

3.3.1 METEOROLOGY

3.3.1.1 General

Meteorological data for Ellerslie were summarized monthly for May through September, 1983 and are presented with longterm data from the Edmonton International and Industrial Airports in Table 3. Ellerslie had higher than normal precipitation during June and lower precipitation during the other months, higher temperatures during August, less sunshine during July and more during August, and lower windspeeds over all.

3.3.1.2 Precipitation

Daily precipitation for the study period is shown in Figure 5. Heavy rain during the 10 day period June 18 through June 27 accounted for 53% (152 mm) of the total precipitation for the study period (May 26 to September 8).

TABLE 3. Summary of meteorological data, 1983

	MONTH				
	MAY	**JUNE**	**JULY**	**AUGUST**	**SEPT**
Temperature (°C)[1]					
Mean daily	11(*10*)	14(*13*)	16(*16*)	17(*14*)	8(*10*)
Mean daily max.	17(*17*)	20(*20*)	22(*23*)	24(*21*)	15(*17*)
Mean daily min.	3(*2*)	8(*6*)	11(*9*)	10(*7*)	3(*3*)
Precipitation (mm)[1]					
Mean precip.	6(*35*)	187(*76*)	78(*99*)	10(*62*)	35(*44*)
No. days precip.	10(*9*)	16(*12*)	21(*13*)	9(*11*)	18(*10*)
Wind (km per day)[2]	10(*17*)	10(*16*)	9(*14*)	6(*14*)	9(*15*)
Bright sunshine (hours)[3]	267(*267*)	228(*251*)	237(*305*)	331(*268*)	165(*186*)

---

[1] Information in brackets from Edmonton International Airport, 1941-1970 (Crown and Greenlee, 1978).

[2] Information in brackets from Edmonton Industrial Airport, 1938-68 (Heapy, 1971).

[3] Information in brackets from Edmonton Industrial Airport, 1930-60 (Heapy, 1971)

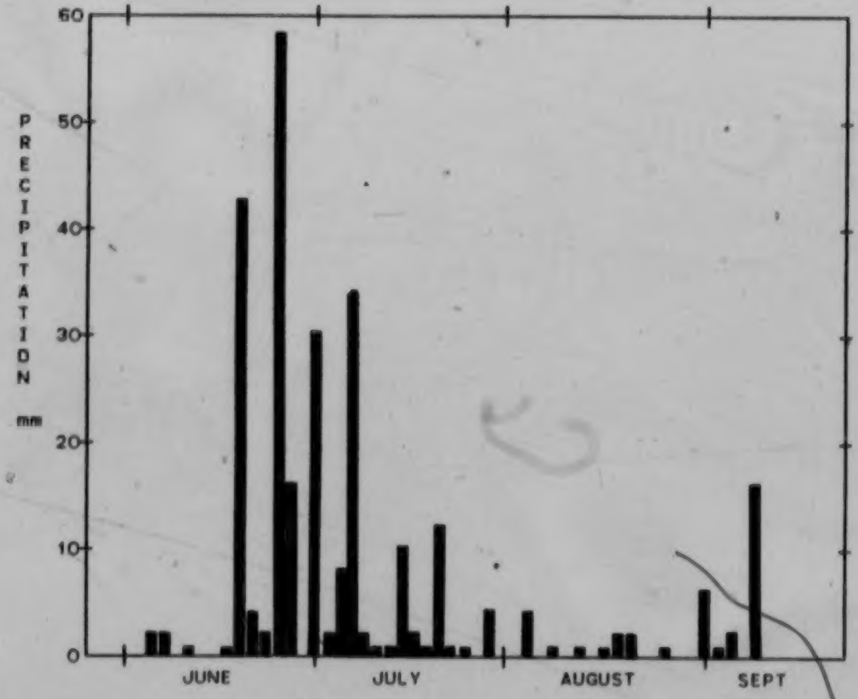


*FIGURE 5. DAILY PRECIPITATION

July had a near normal rainfall of 78.2 mm, as compared with the average of 71 mm (Table 3) but August was quite dry (9.6 mm) as compared with the longterm average of 72 mm. Approximately 46% (133 mm) of the total precipitation for the study period occurred during three rainstorms, on June 18, 24, and July 7, 1983. Precipitation was not as Toogood (1963) had described it for the Edmonton region, 'low in intensity and well distributed'.

Based on infiltration rates measured by Verma (1968) at Ellerslie, runoff from the the Malmo soil is not likely to occur given recorded one-hour intensities for the Edmonton area (Verma and Toogood, 1969). Using ponded infiltration on cultivated soils, rates of approximately 6 cm h^{-1} at 1 hour, and a steady-state rate of 5 cm h^{-1} were recorded. However, as raindrop impact can significantly reduce permeability, the heavy rains in June and July could have resulted in some runoff, especially as very little ground cover existed. After these rainstorms freshly tilled soil was compacted and some micro-rills were observed along wheel tracks.

3.3.1.3 Potential Evapotranspiration

Monthly PET estimates as calculated by the three methods are given in Table 4. Generally the Pan and the Baier-Robertson methods gave approximately equal values, both higher than the Penman estimates. The Penman estimates were similar to those obtained by Verma (1968) and Heapy (1971) when similar albedos were used, except for August

TABLE 4. Potential evapotranspiration

	MAY	JUNE	JULY	AUG	SEPT	TOTAL
	Evapotranspiration (mm)					
Penman						
albedo = 0.25	103	105	128	123	55	391
Verma (1968)[1]	*104*	*114*	*117*	*84*	*46*	*465*
albedo = 0.12	121	123	151	144	64	603
Heapy (1971)[2]	*126*	*139*	*146*	*111*	*72*	*594*
Class "A" Pan						
Ellerslie, 1983	161	159	148	150	84	702
Heapy (1971)[3]	*203*	*206*	*201*	*158*	*91*	*859*
Baier-Robertson[4]	156	147	155	162	96	716

1 Verma (1968) used data from Edmonton Industrial Airport 1955-67 and an albedo of 0.25 to represent a green crop,

2 Heapy (1971) used data from Edmonton Industrial Airport 1959-68 and a variable albedo depending upon crop stage with bare soil before seeding set at 0.12

3 Class "A" evaporimeter data from Edmonton International Airport averaged for years 1968-69 (Heapy, 1971).

4 Calculated from multi-correlation equation by Baier and Robertson (1965) which had higher temperatures. The mean monthly PET increased gradually from May to July and decreased sharply after August.

Laycock (1967) estimated the PET for the Edmonton area using the Thornthwaite method as 508 - 559 mm. Verma (1968) indicated that this was an overestimation for the area as compared to the estimate obtained using Penman's equation; however, the estimate for 1983 (603 mm) using the Penman equation and the parameters used by Heapy (1971) were higher than Laycock's (Table 4).

The Penman method is accepted as being a reliable technique of calculating PET (Hillel, 1980b) as it is physically based. Methods based upon atmometer observations (Baier and Robertson, 1965)) are considered difficult to interpret (Gray, 1970) and have been reported as being over sensitive to wind. In comparing the relative importance of the three major factors involved in evaporation (net radiation, humidity and wind), Mukammal and Bruce (1960; cited in Gray, 1970) found that the proportioning of components for the pan was 80:60:14, and for the Bellani Plate 41:7:52. Sonmor (1963; cited in Gray, 1970) calculated seasonal coefficients to be applied to various evaporimeters to estimate the consumptive use of crops for maximum yield in southern Alberta. For barley, the coefficient for the Black Bellani Plate (cm/cc) is 0.00762 and the coefficient for Class 'A' Pan is 0.66. Nicholaichuk (1964; cited in Gray, 1970) found the Penman method to be applicable for alfalfa growing in soil at potentials higher than -700 kPa for a 14 day period. He used an albedo of 0.25 and found the consumptive use to be equal to 0.95 PET. Gray (1970) suggested that a coefficient of 0.80 be used for crop estimates for May through August.

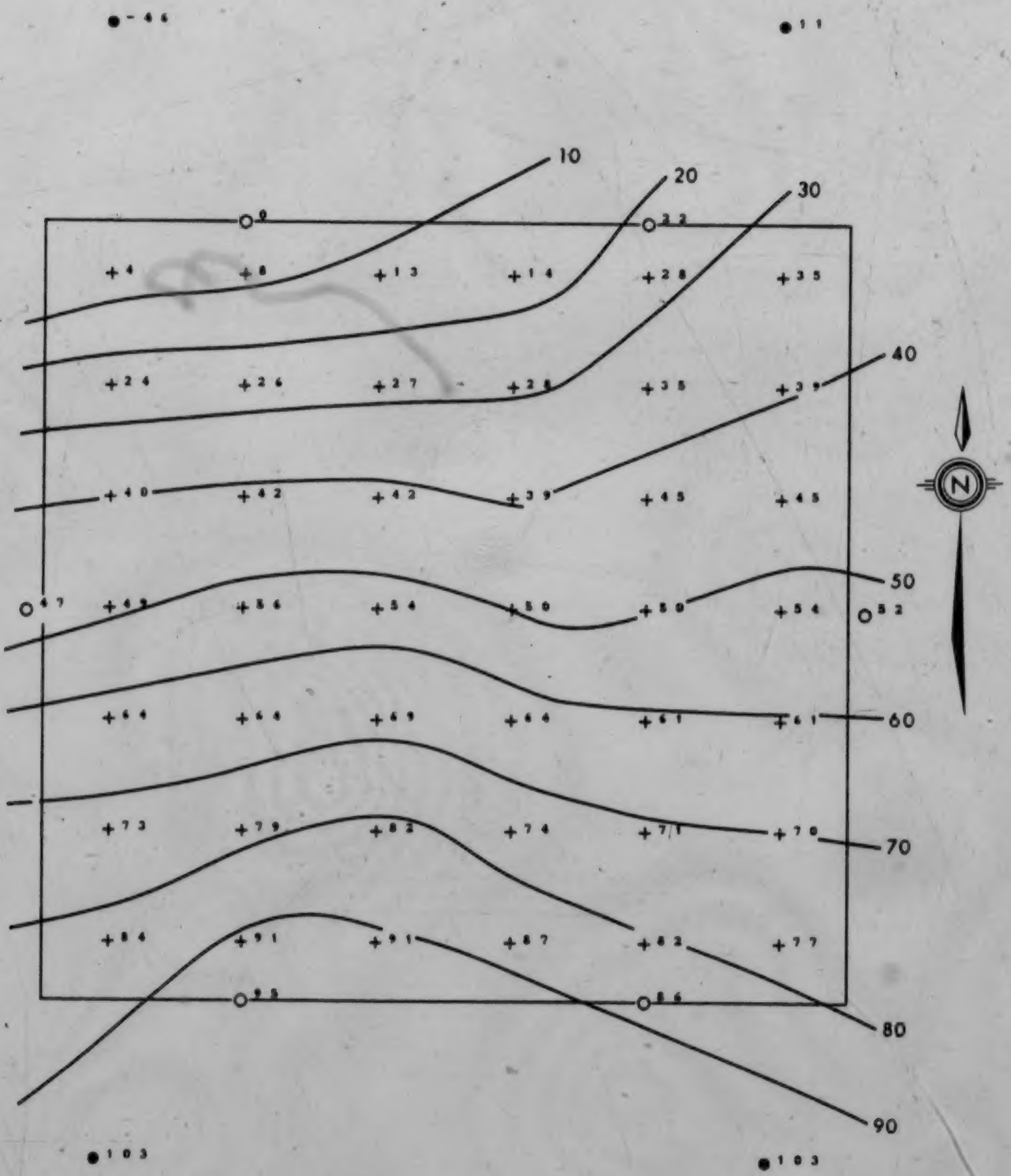
Estimates of AET used in this study to calculate soil water balance were calculated using the Penman method. An albedo of 0.12 was used for bare soil. Barley was assigned albedo values ranging from 0.13 to 0.18, depending upon its stage of growth.

3.3.2 SITE TOPOGRAPHY

Surface elevations of the neutron access tube and water well locations are displayed in Figure 6. The site had approximately a 2% slope to the north with the lowest elevation in the northwest corner.

3.3.3 CROP GROWTH

Due to dry conditions after seeding very little (less than 10%) of the seed germinated until late June when there was sufficient rain for complete germination. The heavy rains in early July initially resulted in poor growth; however, the warm dry weather soon caused excellent growth (Figure 7). The crop was harvested on September 10 in one m^2 samples around each access tube for a total of 21 samples. The resulting total weight of the crop (straw and grain) was 3.24 (±0.53 standard deviations) t ha^{-1}.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+ access tube location
o water well location
• arbitrary elevation location
²³ elevation in cm above lowest point in site
～ 10 cm contour lines

Figure 6. Surface elevations of study plots

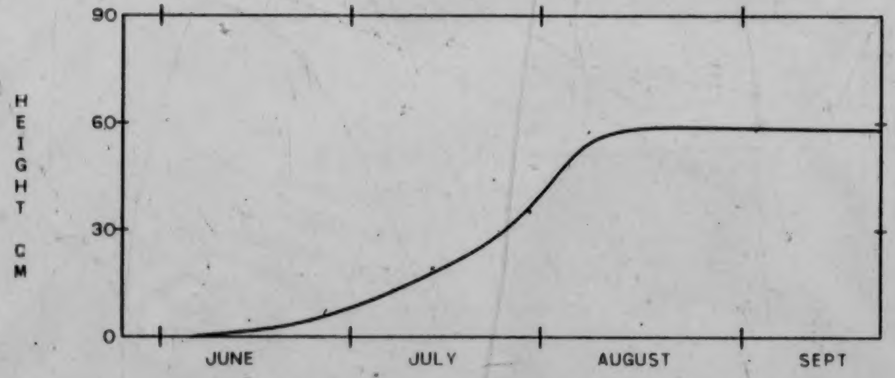

FIGURE 7. BARLEY HEIGHT VERSUS TIME

3.3.4 PHYSICAL PROPERTIES

Bulk density in the 0 to 10 cm layer was quite low (0.98 ±0.11 Mg m^{-3}) due to tillage. There was a steady increase to 1.60 ±0.15) Mg m^{-3} at the 40 to 60 cm depth (Figure 8). The highest average bulk density of 1.69 ±0.01 Mg m^{-3} occurred at the 80 to 100 cm depth.

Soil texture ranged from a loam in the top 20 cm to a clay loam for the rest of the profile. Clay and sand contents increased slightly with depth (Figure 8). Clay increased from an average of 26% (±3.9%) for the uppermost 20 cm to 32% (±6.4) for the 80 to 100+cm depth. Sand increased from a low of 33.6% (±3.8%), for a depth of 0 - 20 cm, to 40.2% (±7.7%) at 40 to 60 cm. Numerous sand layers, approximately 1 to 4 cm thick, were observed below 60 cm. These layers were discontinuous from core to core. As the particle size was determined for samples 20 cm in length, the presence of these sand layers was often masked.

Mass moisture contents for specific pressures generally decreased with depth (Figure 8). The largest decrease occurred between the surface 20 cm and the 20 to 40 cm sample layer. This was possibly due to higher organic matter contents of the surface 20 cm.

Field measurements from the cores indicated that the average depth of the Ah horizon was 33 ±11.5 cm.

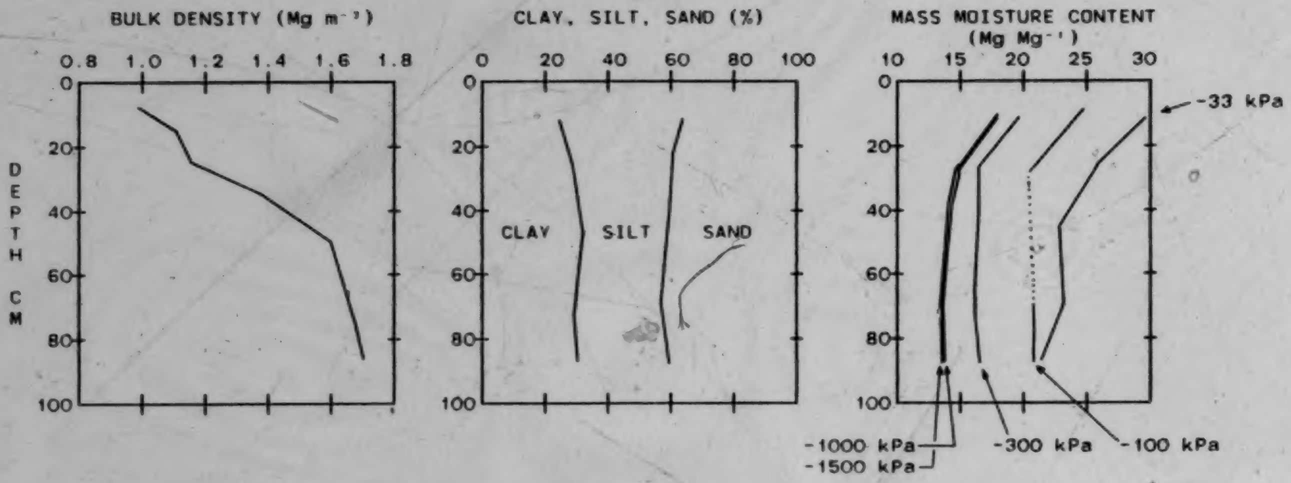

FIGURE 8. BULK DENSITY, TEXTURE, AND MOISTURE RELEASE CHARACTERISTICS

3.3.5 SOIL WATER PROPERTIES

3.3.4.1 Water Table Wells

The water table depth showed a constant and steady decline during the entire measurement period (Figures 9 and 10). The average recorded daily rate of fall, as estimated from regression slopes, was 1.38 (±0.27) cm day^{-1}. During a brief period following establishment of the wells (July 18 to 22), the rate of fall was very small; this being attributed to an equilibration period after installation. Well 3 became clogged with mud after Aug 18.

Initially the water table for the higher ground elevations (sites 4 and 5) was closer to the ground surface (Figure 9) than the lower elevations (sites 1 and 2). The rate of fall for sites 4 and 5 was less (1.19 cm day^{-1}) than sites 1 and 2 (1.71 cm day^{-1}). The water table level also roughly paralleled the surface elevation as indicated when the water levels were measured from a reference plane (Figure 10). The higher the ground surface elevation, the higher the water table relative to the reference plane. The reference plane was established by setting the lowest measured ground surface point to zero. Well sites 3 and 6 were approximately at the same elevation (47 and 52 cm).

Observations by Sanborn (1981) and further observations at this site in June, 1984 (water table at 3.35 m below the ground surface), indicated that the occurrence of a water table at these depths was not common. The water table observed was likely perched due to slowly permeable

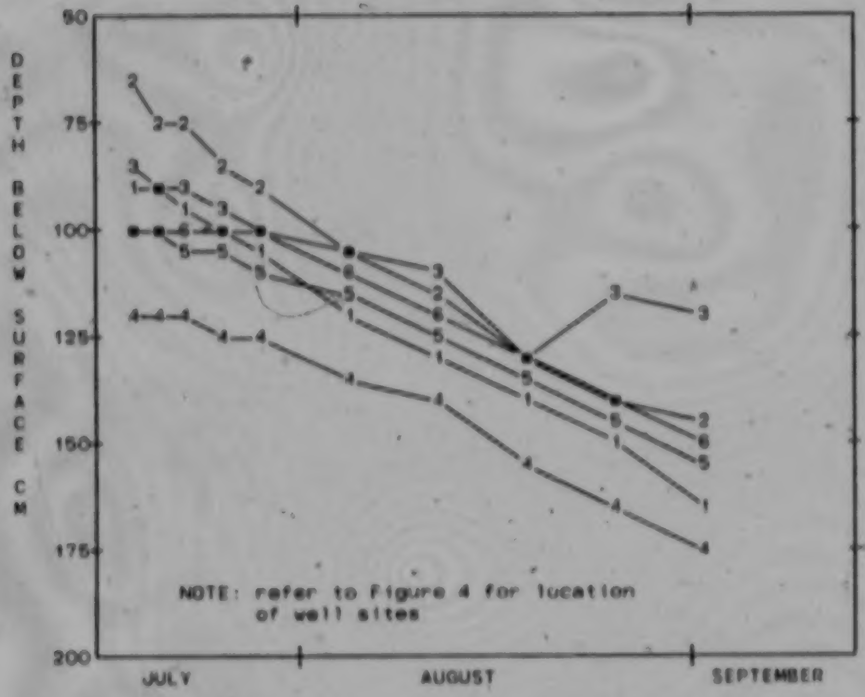

FIGURE 9. DEPTH OF WATER TABLE

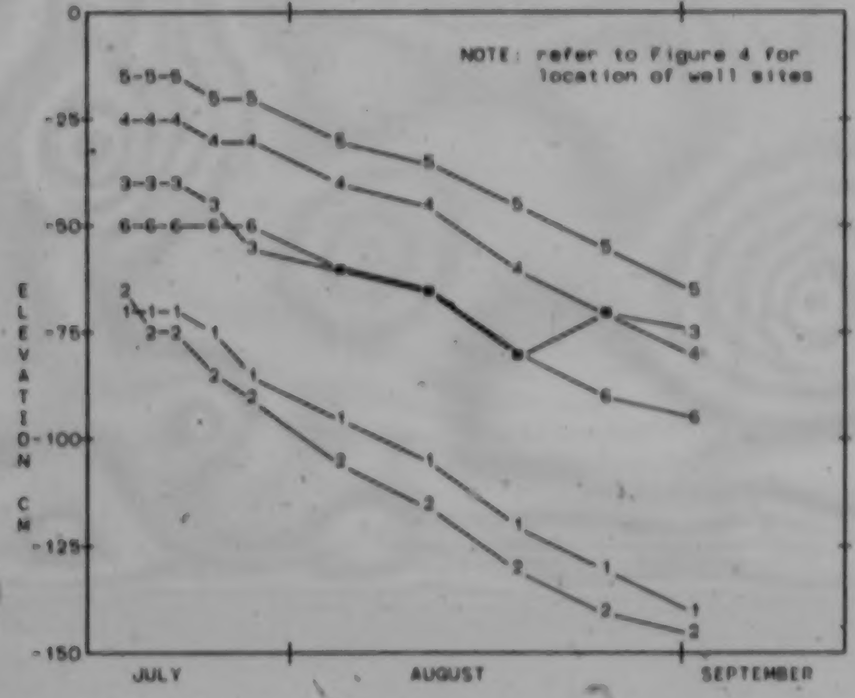

FIGURE 10. ELEVATION OF WATER TABLE

conditions and the unseasonably high rainfall that occurred during the end of June and beginning of July, 1983. The shallow depth of the water table and the fine texture of the soil undoubtedly resulted in the capillary fringe contributing water to evaporation and transpiration. Water tables at 1.8 m beneath the ground surface can contribute up to 25 to 30% of the actual evapotranspiration to a barley crop in a clay loam soil (Read and Pohjakas, 1981).

3.3.4.2 Field Soil Water Potentials

Soil-water potential decreased for all depths at both treatments with time (Figures 11 and 12). The soil water potential for the barley plots decreased faster and to greater depths than that for the fallow plots. Potentials measured at 75 and 90 cm remained quite high (greater than -33 kPa) during the growing season due to the presence of a high water table.

Generally very little change in soil water potential occurred until after July 11 due to poor barley growth, heavy rains and high water table conditions. Readings taken on June 24 and June 28, showed positive potentials at depths of 45 cm and below. Readings at the 30 cm depth generally remained greater than -5 kPa until July 4th to 8th when the barley began vigorous growth. Soil water potential at this depth did not reach values less than -10 kPa until after July 14th, and -33 kPa until August 4th for the barley plots and August 11 for the fallow plots. Soil water potentials

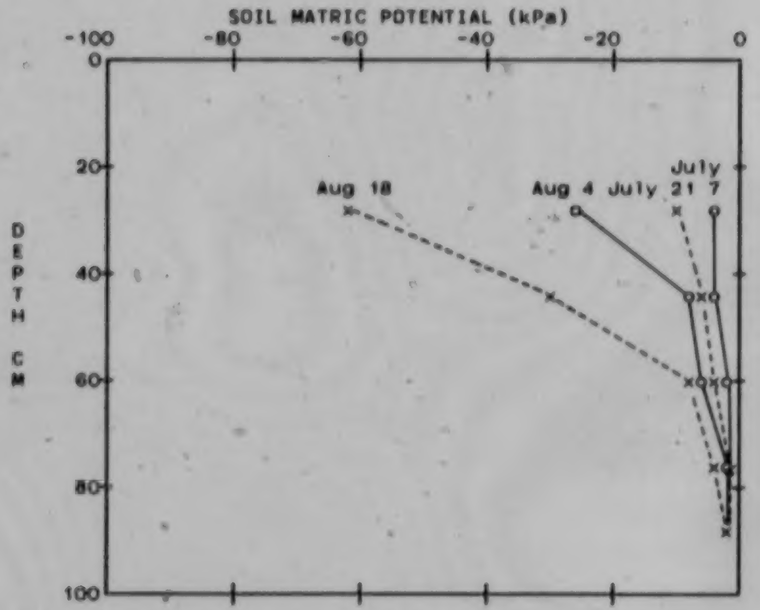

FIGURE 11. PROFILE MATRIC POTENTIAL WITH TIME FOR FALLOW PLOTS

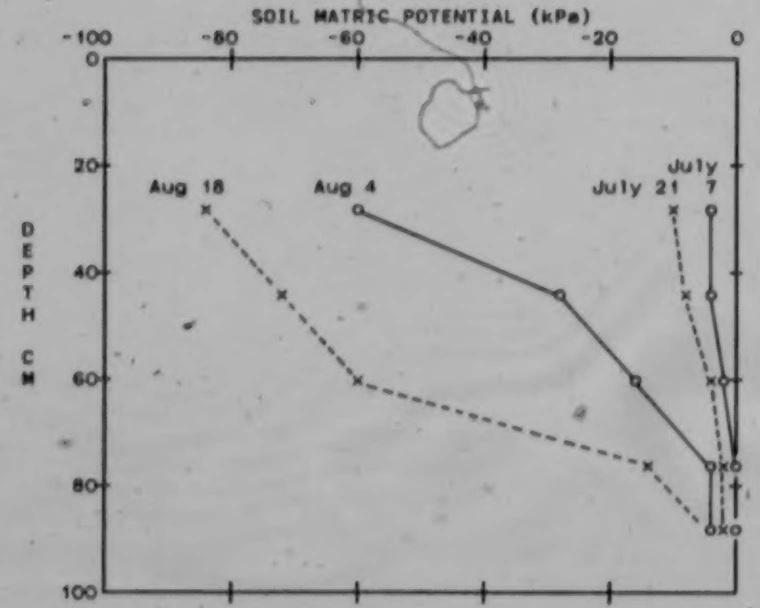

FIGURE 12. PROFILE MATRIC POTENTIAL WITH TIME FOR BARLEY PLOTS

for the fallow plots at and below 60 cm remained greater than -10 kPa until August 25. Some of the 90 cm tensiometers for both treatments were suspected as giving erroneous readings in late August due to the development of large soil cracks near the tensiometers. Readings near the tensiometer limits (-60 to -80 kPa) became quite variable, due to air entering the water column and due to soil drying.

Hydraulic gradients for fallow plots on August 4 (day 70) averaged 11.9 cm cm^{-1} between 30 and 45 cm, whereas barley plots averaged 20.4 cm cm^{-1}. The hydraulic gradient during August 4 between the 30 to 90 cm depths averaged 4.2 and 9.7 cm cm^{-1} for the fallow and barley plots respectively.

The positive potentials measured by the tensiometers in late June and early July indicated a high water table. Tensiometers have been successfully used (Richards *et al.*, 1973) to estimate water table depth, by assuming that zero potential exists at the water table surface, positive potential below and negative above. Tensiometer readings on June 24 did not indicate the presence of any water table to a depth of 90 cm. Between June 24 and June 27 there was 72.8 mm of precipitation. The water table depths for June 27 as inferred from the tensiometer readings were between 37 and 39 cm. By June 30 the only water table recorded was in plot E at 85.5 cm. It was thus assumed that the water tables recorded on June 27 were actually due to saturated but still draining conditions. Although very little rain occurred between June 29 and July 4, the water table was recorded at the 60 cm depth on July 4. By July 14 only one plot had a water table within measurable range at 85.2 cm. On July 18 no water tables were within tensiometer range; however, well #3 recorded the water table level at 87 cm.

3.3.4.3 Moisture Content

Volumetric soil moisture increased with the high amounts of precipitation in late June and early July and decreased then due to increasing PET, crop growth and the water table dropping (Figures 13 and 14). The heavy rainstorms in late June and early July caused a rapid increase in soil moisture, with the 15 cm depth having the highest moisture content. By late July to early August the barley plots were much drier than the fallow plots. Moisture contents at 90 and 105 cm for both plots remained high due to the shallow water tables. As the water table was perched, some of the gradual decline in moisture contents at the deeper depths was likely due to the fall in water table. The lower moisture contents at the 90 and 105 cm depths in the barley plots, relative to the fallow plots, can be assumed to be due to greater hydraulic gradients and/or to root extraction.

Changes in moisture content between barley and fallow plots became apparent after July 28 (Figures 15 and 16). For the fallow plots large changes in moisture content (>0.05 m^3 m^{-3}), between dates occurred only at depths <45 cm. The

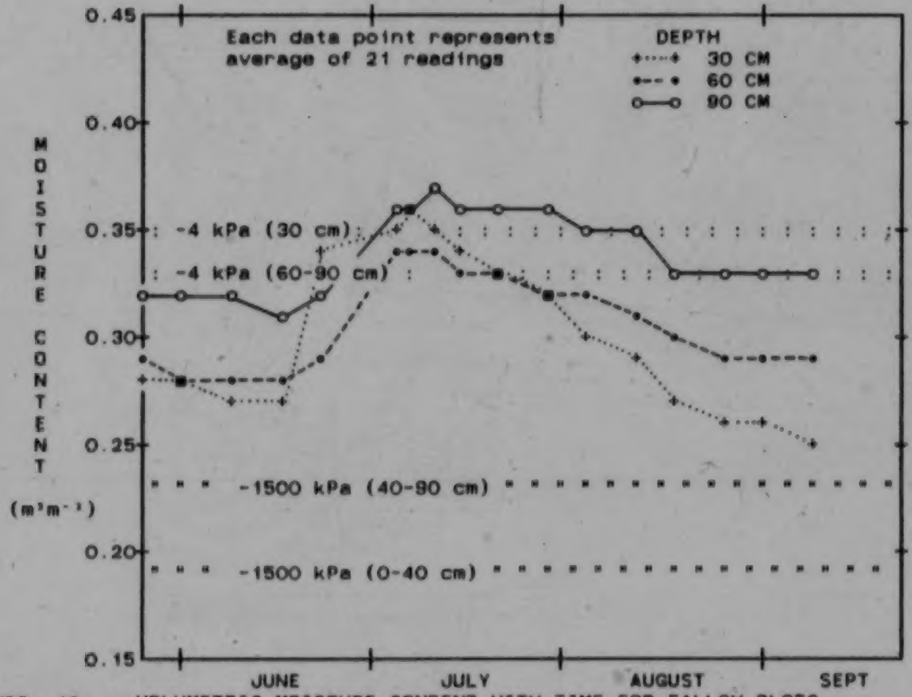

FIGURE 13. VOLUMETRIC MOISTURE CONTENT WITH TIME FOR FALLOW PLOTS

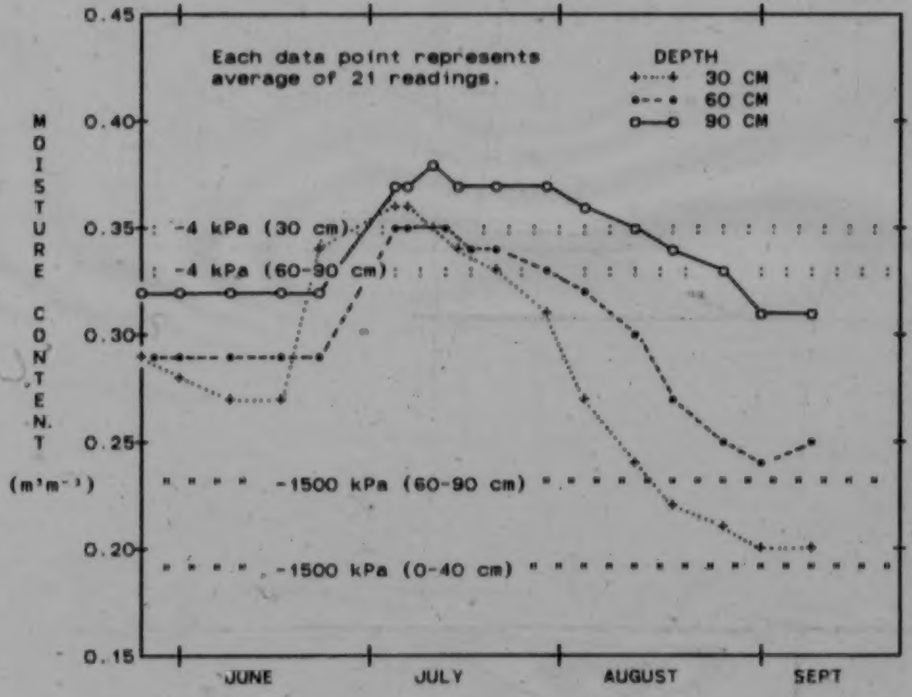

FIGURE 14. VOLUMETRIC MOISTURE CONTENT WITH TIME FOR BARLEY PLOTS

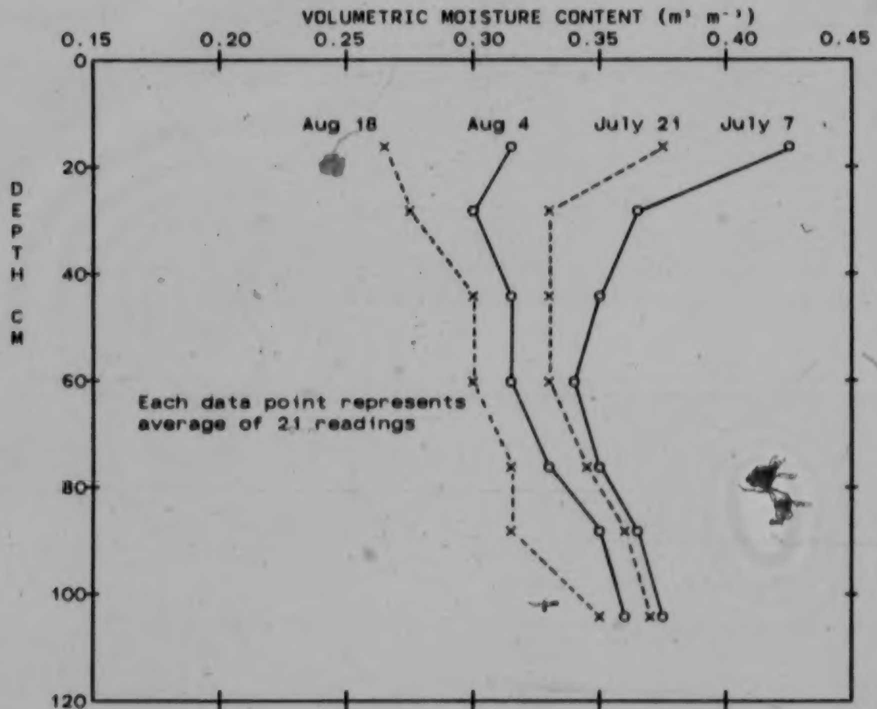

FIGURE 15. PROFILE MOISTURE DISTRIBUTION WITH TIME FOR FALLOW PLOTS

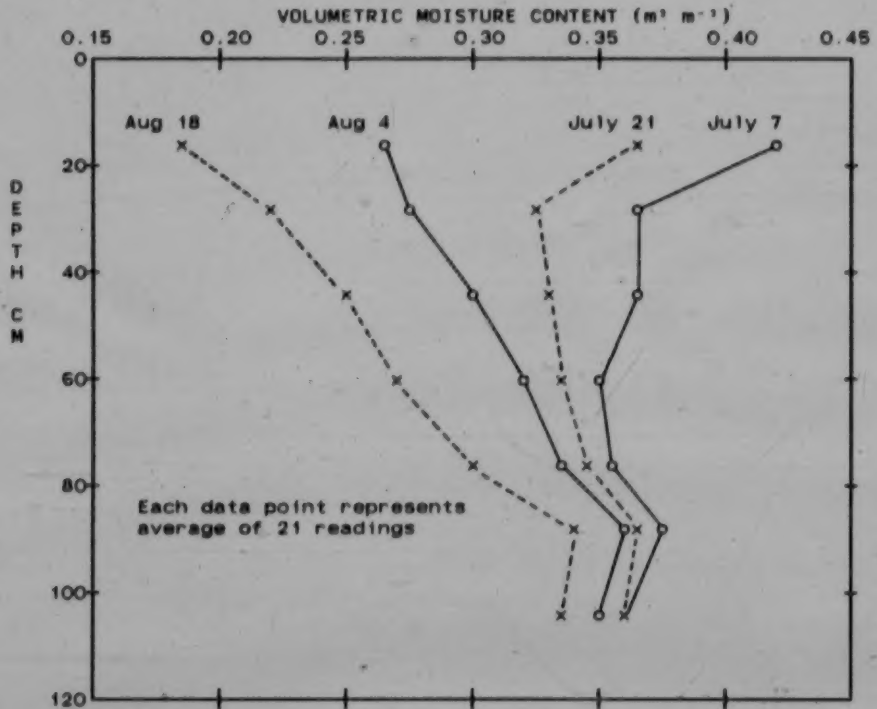

FIGURE 16. PROFILE MOISTURE DISTRIBUTION WITH TIME FOR BARLEY PLOTS

barley plots, however, had large changes in moisture content to depths greater than 75 cm.

3.3.4.4 Water Holding Capacity

Water holding capacities as determined from tensiometers and from pressure plates are reported in Table 5. Field soil potentials of -4 kPa were chosen as representative of field capacity based on results by Russell (1961), Webster and Beckett (1972) and Parkes and O'Callaghan (1980). For the pressure plates -33 kPa was used to represent field capacity as this is the value commonly used (Richards, 1965). Permanent wilting point was only measured using the pressure plates. Gravimetric moisture contents from the pressure plate analysis was converted to a volume basis using field bulk densities. Moisture contents (m^3 m^{-3}) at -4 and -33 kPa for all the tensiometers were determined by plotting the characteristic curve and interpolating the values (Figure 17) .

The water holding capacity (129 mm m^{-1}) as determined from the tensiometers using -4 kPa to represent field capacity, was slightly lower than that measured using the pressure plates. If -33 kPa was used as representative of field capacity for the tensiometers, the water holding capacity would only be 87 mm m^{-1} (Table 5). The reason for dissimilar field capacities between the tensiometers and the pressure plates could be due to the extrapolation of moisture contents from the pressure plates to field

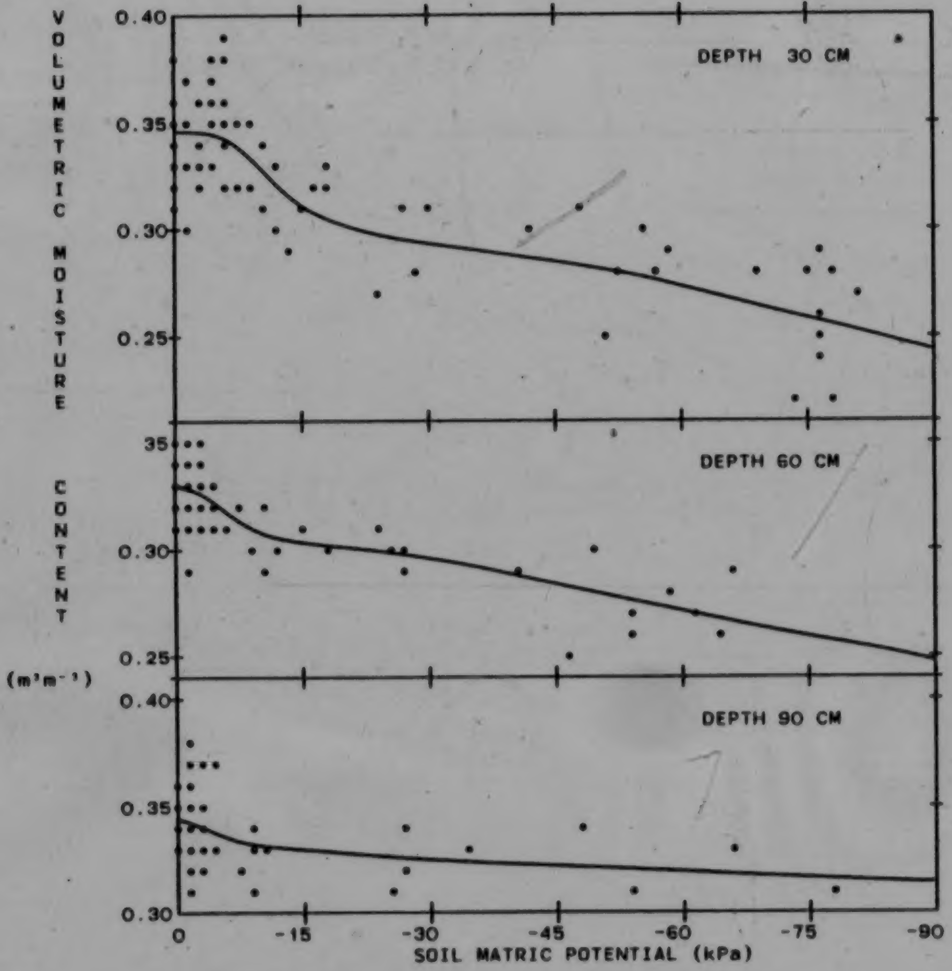

FIGURE 17. MOISTURE CHARACTERISTIC CURVES FOR TENSIOMETERS AT 30, 60, AND 90 CM

TABLE 5. Field capacity and permanent wilting point volumetric moisture contents

	FIELD CAPACITY (m^3m^{-3})			PWP (m^3m^{-3})
DEPTH	**TENSIOMETER**		**Pr Pl**	**Pr Pl**
(cm)	-4 kPa	-33 kPa	-33 kPa	-1500 kPa
0- 20			0.32	0.19
20- 40			0.32	0.19
30	0.35	0.29		*0.19**
40- 60			0.36	0.22
45	0.34	0.29		*0.22**
60	0.33	0.29		*0.22**
60- 80			0.39	0.22
75	0.33	0.29		*0.22**
80-100			0.35	0.23
90	0.33	0.32		*0.23**

* values at these depths interpolated from sampled depths.

Pr Pl - pressure plates

situations. The grinding and sieving pretreatment altered the pore size distribution. This resulted in higher field capacities than those obtained from the tensiometers, especially for depths below 60 cm. This effect is likely minimized at the lower potentials of -1500 kPa, as moisture content in this range is largely determined by texture and not structure (Rode, 1969; Gumma'a, 1978).

Water holding capacities reported by Verma (1968) and Heapy (1971) for the Malmo soils at Ellerslie were 184 mm and 204 mm respectively for the top 1.0 m of soil, considerably higher than that reported for this study (129 and 138 mm). Both studies used pressure plate values (33 and 1500 kPa) and bulk density for determining water holding capacities. Field capacities were much higher, especially for the depths below 30 cm (Verma, 60-80 cm, 42.5 $m^3 m^{-3}$; Heapy, 60-90 cm, 45.3 $m^3 m^{-3}$) and permanent wilting points were equivalent. Discrepencies are likely due to textural differences, as the soils in this study had higher sand contents (10 to 15% higher) and lower clay contents (4 to 10%) than those reported by Verma (1968) and Heapy (1971).

3.3.6 SOIL WATER BALANCE

3.3.6.1 Changes in Soil Moisture (ΔS)

Changes in total soil moisture in the top 100 cm under barley and fallow conditions were similar until mid July, at which time changes under barley exceeded those of fallow (Figure 18). Gains in soil moisture

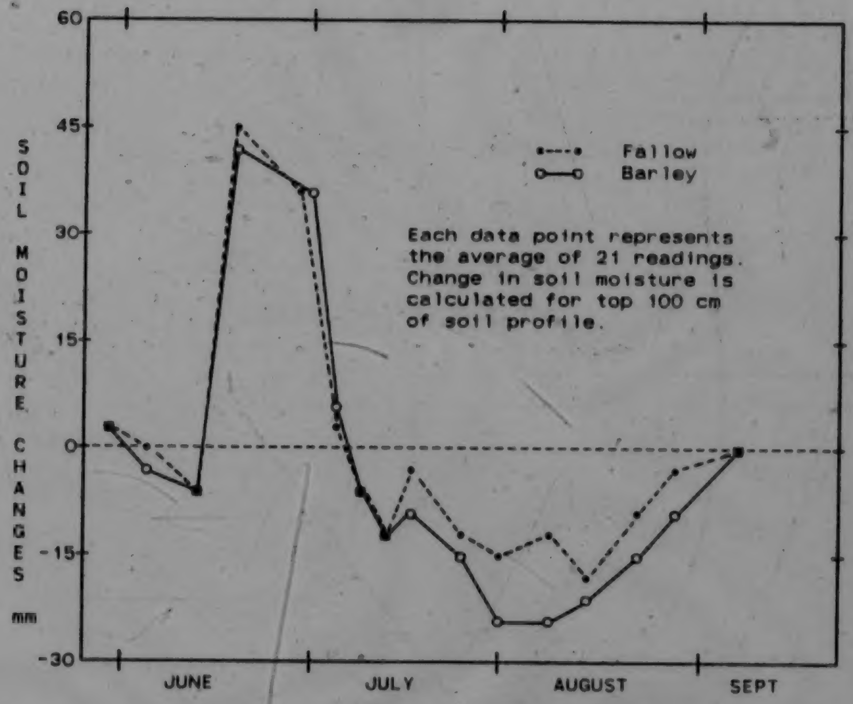

FIGURE 18. SOIL MOISTURE CHANGES FOR BARLEY AND FALLOW

(+ΔS) were due to precipitation and losses (-ΔS) were due to drainage and evapotranspiration.

Changes in soil moisture (ΔS) for the top 1 m, precipitation (P), and potential evapotranspiration (PET) between consecutive recording dates are presented in Table 6. If the ΔS for these periods is subtracted from the amount of precipitation that occurred during the same time period, the difference is the amount that occurred as runoff, drainage, and/or evapotranspiration. These amounts are shown for both fallow and barley conditions under the column labelled 'P - ΔS'. Generally, except during the rainy period from June 16 to July 7 and for some periods of high moisture loss from the barley plots during July and August, PET exceeds $P - \Delta S$.

From Figure 18 and from the data in Table 6 the moisture balance for the study period may be divided into three periods:

1. May 26 to June 16: little rainfall, no growth, and little change in ΔS.
2. June 16 to July 7: high rainfall, no growth, and large positive changes in ΔS.
3. July 7 to Sept 8: little rainfall, vigorous barley growth, and large negative changes in ΔS.

As the moisture content was below field capacity during the first period, any water loss (P - ΔS) may be attributed to Et. During the second period and the first

TABLE 6. Determination of moisture balance

DATE	ΔS (mm)		P (mm)	$P - \Delta S$ (mm)		*PET*[1] (mm)	
	Fal	Bar		Fal	Bar	Fal	Bar
May 26							
June 1	-5	-5	0	5	5	25	25
June 9	1	-2	4	3	6	30	30
June 16	-6	-4	tr	6	4	31	31
June 23	44	44	64	20	20	18	18
July 4	35	37	121	86	84	28	28
July 7	4	5	42	38	37	13	13
July 11	-6	-5	4	10	9	17	17
July 14	-11	-12	tr	11	12	13	13
July 21	-4	-8	25	29	33	24	20
July 28	-11	-17	1	12	18	28	27
Aug 4	-15	-25	6	21	31	34	32
Aug 11	-11	-26	1	12	27	32	30
Aug 18	-17	-23	3	20	26	23	22
Aug 25*	-11	-16	3	14	19	26	24
Sept 1	-4	-8	tr	4	8	20	18
Sept 8	0	1	10	10	9	14	14
Total	-17	-64.0	284	301	348	376	362

[1] *PET* calculated from Penman equations:
albedo for bare soil 0.12,
albedo for early barley growth 0.13 to 0.18 (July 21 to August 4)
albedo for full green barley cover 0.18.
Crop coefficient for conversion of Penman estimate to barley Et 0.8 (Gray, 1970).
Bare soil coefficient for conversion of Penman estimate 0.8 (Hartmann *et al.*, 1980)

* corrected due to change in neutron probes.

portion of the third period when the moisture contents were above field capacity, some of the moisture lost was also due to drainage. Due to the vigorous barley growth, higher *PET* rates during the third period, and moisture contents below field capacity, most of the water loss was probably due to *Et*. Changes in soil water status for each of these periods will be discussed in greater detail in the following subsections.

3.3.6.2 Drainage (*D*) and Evapotranspiration (*Et*)

General Approximation

Assuming no water losses by runoff or interflow, and no water gains from groundwater, a simple estimation of water loss due to both evapotranspiration and drainage for the study period (May 26 to September 8) for a 1 meter depth may be calculated from the data in Table 6 using the following simplified version of Equation (1):

$$P - \Delta S = Et + D \qquad (16)$$

Barley: 284 - (-64) = 348 mm

Fallow: 284 - (-17) = 301 mm

The net change in soil moisture for the season was negative, with barley having the greater loss of 64 mm. Thus barring any water addition or loss due to runoff or interflow, the total amount of water lost due to evapotranspiration and drainage was 348 mm for barley and 301 mm for fallow. For this general calculation, which includes the entire time period, any contributions from the shallow groundwater table does not affect these totals of 348 and 301 mm as the water table formed during this period from the heavy rains during June 16 to July 7.

Approximation Using The Field Capacity Concept

The drainage and evapotranspiration components were separated with the use of the following assumptions:

1. drainage occurred only when the moisture content was above field capacity, as measured at -4 kPa by the tensiometers;
2. when soil moisture was below field capacity any losses were due only to evapotranspiration; and
3. moisture loss due to evaporation was equivalent to potential evaporation as measured by the Penman method only while the soil moisture for the profile was above field capacity.

The evapotranspiration and the drainage components for fallow and barley conditions are listed in Table 7 in mm of water for the top 1m of profile for the individual time periods along with ΔS.

For period 1 (May 26 to June 16) all losses were attributed to *Et* as the moisture content was below field capacity. Total *Et* for this period was:

TABLE 7. Evapotranspiration and drainage components: Field capacity method

DATE		FALLOW			BARLEY	
	Δ*S* (mm)	*Et* (mm)	*D* (mm)	Δ*S* (mm)	*Et* (mm)	*D* (mm)
May 26						
June 1	-5	5		-5	5	
June 9	1	3		-2	6	
June 16	-6	6		-4	4	
June 23	44	18	2	44	18	2
July 4	35	28	58	37	28	56
July 7	4	13	25	5	13	24
July 11	-6	10		-5	8	1
July 14	-11	7	4	-12	8	4
July 21	-4	28	1	-8	31	2
July 28	-11	10	2	-17	16	2
Aug 4	-15	20	1	-25	29	2
Aug 11	-11	11	1	-26	26	1
Aug 18	-17	20		-23	26	
Aug 25	-11	14		-16	19	
Sept 1	-4	4		-8	8	
Sept 8	0	10		1	9	
Total	-17	207	94	-64	254	94

	Et	D
Fallow	14 mm	0 mm
Barley	15 mm	0 mm

During the second period (June 16 to July 7) moisture contents throughout the entire profile were above field capacity due to the large amount of rainfall. Consequently drainage was assumed to have occurred. Precipitation not accounted for by the positive changes in ΔS ($P - \Delta S$ in Table 6) was assumed to be lost due to Et which was set as equivalent to the PET in Table 6 with the remainder equal to the drainage (Table 7), that is,

$$P - \Delta S - PET = D \tag{17}$$

The Et and D components so calculated for the second period are listed in Table 7 for the individual measuring times. The total amounts for period 2 are listed below.

	Et	D
Fallow	59 mm	85 mm
Barley	59 mm	82 mm

After July 7 drainage and evapotranspiration were partitioned according to the depth of occurrence of field capacity as represented by -4 kPa. Water loss from depths at which the moisture content was above field capacity was assumed to be due to drainage and water loss from depths at which the moisture content was below field capacity was assumed to be due to evapotranspiration. The depths and dates at which potential decreased below -4 kPa are represented in Table 8 by the letters "F" and "B". Any precipitation during this period was assumed to have been lost due to evapotranspiration. *Et* and drainage so separated for the period from July 7 to September 8 are listed in Table 7 and totalled for this period below:

	Et	*D*
Fallow	134 mm	9 mm
Barley	180 mm	12 mm

The total amounts of moisture lost due to evapotranspiration and drainage during the entire study period (May 26 to September 8) as summed from Table 7 are:

	Et	*D*	Total
Fallow	207 mm	94 mm	301 mm
Barley	254 mm	94 mm	348 mm

Barley utilized only 47 mm more moisture by *Et* than did the fallow. The *Et* estimates above are likely smaller than *Et* losses that actually occurred due to possible contributions from the shallow water table. As the drop in water table during the season cannot be separated into the components of deeper drainage or upward flux, an estimate of contribution of the water table to *Et* cannot be given.

Approximation by Gradient Method (Barley only)

Graphical examination of changes in soil moisture content versus time (Figures 13 and 14), according to the methodology outlined by McGowan and Williams (1980a), (see Section 3.1.7, page 40 of this report) did not display sufficient resolution to discern changes in slope attributable to evapotranspiration. Consequently a modification to the method was used. Using the fallow plots as a control, root extraction from the barley plots was assumed to start where and when the soil moisture between plots began to differ significantly.

Testing of significant differences was performed with the Wilcox non-parametric test for independent pairs (Steel and Torrie, 1978). A non-parametric test was used in place of a parametric test (eg. paired t-test) because the data could not be assumed to be normally distributed. Two sets of data were tested; moisture contents for barley and fallow; and changes in moisture content for consecutive dates for barley and fallow. With this method the evapotranspiration and drainage components may be separated only for the barley plots during active growth which occurred during the third period (July 7 to September 8).

Separation of the Et and D components before barley growth was made with the same techniques and assumptions as described for the first two periods in the field capacity section.

The results of the Wilcox test of significance for the individual depths and times of measurement are displayed in Table 8 for the moisture contents and in Table 9 for the changes in moisture contents. Both sets of tests show root extraction beginning on the same date, July 21, at the depths of 15 and 45 cm for moisture contents (Table 8) and at the depths of 15, 30, and 45 cm for changes in moisture (Table 9).

The root extraction depths were greater each week and by August 25, both the moisture contents was significantly smaller and the changes in moisture were significantly larger in the barley plots at 90 cm. Because the rainfall was low during this period, the moisture contents remained significantly different for all depths through to September 8 (Table 8). As the roots grew deeper and as the moisture content decreased in the upper parts of the profile (Figure 14), roots extracted the more readily available water in the lower parts of the profile. This is evident from the lack of significant differences occurring for the shallow depths

Table 8. Comparisons between barley and fallow moisture contents (θ)

DATE	DEPTH (cm) 15	30	45	60	75	90
May 26	+	/	/	/	/	/
June 1	/	/	/	/	/	/
June 9	+++	/	/	/	/	/
June 16	+++	/	/	/	/	/
June 23	++	/	/	/	/	/
July 4	/	/	+++	/	/	/
July 7	/	/	+	-	/	/
July 11	/F	/F	/	/	/	/
July 14	/B	/B	/	/	/	/
July 21	++	/	+++FB	/	/	/
July 28	+++	++	+++	/F	/	--
Aug 4	+++	+++	+++	/B	/F	/
Aug 11	+++	+++	+++	++	/B	/F
Aug 18	+++	+++	+++	+++	++	/B
Aug 25	+++	+++	+++	+++	+++	++
Sept 1	+++	+++	+++	+++	+++	+++
Sept 8	+++	+++	+++	+++	+++	+++

Notes:

B - Barley field capacity: cessation of drainage at and after this point

F - Fallow field capacity: cessation of drainage at and after this point

Barley > Fallow Symbol	Fallow > Barley Symbol	Significance
+	-	0.05
++	--	0.02
+++	---	0.01

/ no significant difference

Results are based upon Wilcox's test on independent pairs (two-tailed test).

TABLE 9. Comparisons between barley and fallow changes in soil moisture (ΔS)

DATE	DEPTH (cm) 15	30	45	60	75	90
June 1	+	/	/	/	/	/
June 9	+++	/	/	/	+++	/
June 16	/	/	/	/	/	/
June 23	/	/	/	/	++	/
July 4	/	/	/	/	/	/
July 7	/	/	/	--	/	/
July 11	/	/	/	/	/	/
July 14	/	/	/	/	/	+
July 21	++	++	+++	/	/	/
July 28	+++	+++	/	/	/	-
Aug 4	+++	+++	+++	++	/	/
Aug 11	+++	+++	+++	+++	+++	/
Aug 18	/	++	+++	+++	+++	/
Aug 25	/	/	/	+++	+++	+++
Sept 1	/	/	/	+++	+++	+++
Sept 8	/	/	/	/	/	/

Notes:

Barley > Fallow Symbol	Fallow > Barley Symbol	Significance
+	-	0.05
++	--	0.02
+++	---	0.01

/ no significant difference

Results are based upon Wilcox's test on independent pairs (two-tailed test).

on August 18 and thereafter as shown in Table 9. By September 8 the changes in moisture contents at all depths were not significantly different between the fallow and the barley plots. At this time root extraction and barley growth were assumed to have ceased.

The depths and times at which these significant differences occurred are assumed to demarcate the moisture losses due to drainage from those due to evapotranspiration.

Comparison of Methods and Comments

Evaporation according to the field capacity method began 1 to 2 weeks before that determined by the gradient method for most depths (Table 8). This could be because the field capacity method is more sensitive to losses from evaporation whereas the gradient method measures the onset of transpiration. It is expected that water losses from shallow depths (0-45 cm) would occur from evaporation first before the roots began extraction; however, the 1 to 2 week lag period is consistent to the 90 cm depth. The field capacity method also assumes that the soil profile is homogeneous without any textural discontinuities that can affect the drainage process (Miller, 1973; Gardner, 1979; Hillel, 1980b). This site was characterized by sandy layers at depths below 60 cm, which in one location resulted in a sudden decrease in moisture content at the 60 cm depth between July 28 and August 11 relative to the rest of the site.

Another explanation might be that these methods are not entirely representative of conditions at which drainage ceases and evapotranspiration begins. Perhaps field capacity is better represented by -6 or -10 kPa. The gradient method, as used in this study, is more a measure of when water began to be removed by root extraction. It is very possible and likely that at the shallower depths upward flow as a result of evaporation from the surface had already begun.

The total drainage and evapotranspiration calculated using the two methods are listed below:

Method	plots	*P*	ΔS	*Et*	*D*
Field Capacity	Fallow	284	-17	207	94
	Barley	284	-64	254	94
Gradient (θ)	Barley	284	-64	260	88
(ΔS)	Barley	284	-64	261	87

Despite the one to two week lag that occurred between the field capacity and the gradient depths in discerning the onset of *Et* and the cessation of drainage, the actual difference is only 6 mm for a 2% difference for *Et* and 6mm for a 5% difference for *D*.

Drainage could have also occurred under conditions which would not have been measured by either method.

Initial rapid drainage through structural cracks can take place during and immediately after heavy rainstorms (Quisenberry and Phillips, 1976; Parkes and O'Callaghan, 1980). If this did occur then the current estimate of *D* would be low and the *Et* estimate would be high.

3.3.6.3 Water Table Contributions (*U*)

The amount of soil moisture lost due to evapotranspiration was undoubtedly augmented by groundwater contributions due to the shallow water table. The steady drop in the water table during July and August would then be due to both deeper drainage and to upward flow; however, as the hydraulic conductivity and the drainage conditions of the deeper subsoil are not known, the amount contributed to *Et* cannot be calculated. Studies in other regions indicated that between 30 and 70% of the *Et* for a crop can be contributed by a water table at a depth of 100 to 300 cm (Purvis, 1964; Saini and Ghildyal, 1978; Read and Pohjakas, 1981). The actual amount is dependent upon the depth of the water table, the soil texture, the type of crop, and the hydraulic gradient.

Read and Pohjakas (1981) recorded net seasonal water use for barley grown near Lethbridge, during 1980 and 1981, as varying between 374 and 506 mm for a loamy sand soil and a clay loam soil. Grain yield for the barley crop was correlated to the water use. The gross sample weight for the crop varied between 3.5 and 4.0 t ha^{-1} for 1981. The gross sample weight for the barley crop from this study site was 3.24 (±0.53) t ha^{-1} with an estimated water use (*Et*) of 254 to 261 mm, considerably less than that estimated by Read and Pohjakas (1981). They calculated that the water table, at 185 cm in a clay loam soil, contributed 30% of the total *Et*. If the drop in the water table can be entirely attributed to upward flow, then the maximum amount of *Et* by the barley crop at the Ellerslie study was 348 mm. The drainage component of 94 mm represents 27% of 348 mm.

3.3.6.4 Runoff and Interflow

Although infiltration rates recorded by Verma (1968) at the Ellerslie Research Station on Malmo soil indicate that runoff is not likely to occur given typical rainstorm intensities for the Edmonton area, the occurrence of surface sealing from raindrop impact was not taken into consideration. The presence of micro-rills along tractor tracks within the plots after the occurrence of the heavy rainstorms in late June and early July was an indication that some runoff did occur. The previously loose surface of the tilled fallow plots had also been compacted by the rains. The site was located on a local rise and consequently any runoff that occurred would have resulted in a net loss. Average slope was 1 to 2% to the northwest with the steepest slope at 3.2% in the northwest corner. Because the slopes were slight, runoff is believed to have been very small.

With the slight slope; the near saturated conditions resulting from the heavy rainfalls; the high bulk densities of the deeper depths; the low permeabilities of the deeper subsoil materials, as evident by the perched water table; and the slope of the water table itself, interflow might have occurred during periods of high intensity rainfall. If interflow and runoff did occur then it would be expected that the lower end of the plot would be wetter. During the near saturated conditions (July 4 to July 11) the north edge and the northwest access site were measured as slightly wetter than the rest of the site. This, however, is complicated by moisture and textural variability between the sites and also by the fact that the water table is closer to the surface in the northwest corner.

4. SPATIAL VARIABILITY OF SOIL-WATER

4.1 INTRODUCTION

Soil is continuously variable in space. Proper statistical description of its variability should include more than determination of the mean and dispersion. Soil properties change with distance, consequently adequate description should also consider the spatial variability. Through measurement of how the values change with distance, the relative degree of similarity can be measured. If a property shows greater similarity between neighbouring observations than those further away, it is considered to be spatially dependent. With knowledge of the spatial dependence of a property, future sampling programs can be more efficiently designed, through increasing the precision with fewer samples (McBratney and Webster, 1983).

4.2 EXPLANATION OF STATISTICAL METHODS

4.2.1 MEASUREMENT OF DISPERSION

4.2.1.1 Sampling Precision

A set of collected observations may be summarized into two statistical parameters; the *mean* which describes the typical observation; and the *standard deviation*, which describes the amount of dispersion about the mean. The *precision* of a set of observations refers to the width of the dispersion about the mean (Kempthorne and Allmaras

(1965).

Using the mean and the standard deviation an estimate of the required sample size to achieve a desired precision may be calculated (Cline, 1944; Petersen and Calvin, 1965):

$$n = \frac{t_k^2 s^2}{D^2} \tag{16}$$

where n is the required number of samples; t_k^2 is the estimated value of Student's t at the chosen level of probability, k; s^2 is the estimated variance; and D^2 is the 'specified limit' or the level of precision.

Another expression of variability is the *coefficient of variation* (*CV*), which is also referred to as the relative dispersion or the coefficient of dispersion. The *CV*, defined as the standard deviation divided by the mean, expresses the standard deviation free from units of measurement, enabling the relative dispersion of one soil property to be compared to that of another. Wilding and Drees (1983) caution that when a directly proportional relationship exists between the magnitude of the mean ($\bar{x}$) and s, *CV* is an invalid index. Rao *et al.* (1979) stated that the *CV* does not provide an adequate insight into the nature of the dispersion of the measured population as it does not indicate the degree of normality. Nevertheless, the CV has been a widely used parameter (in soil science) to present a quantitative index of the amount of variability.

4.2.1.2 Normal and Anormal Distributions

Statistical measurements describing the shape of the distribution are the variance, skewness, and kurtosis. As parmetric statistics assume that the population has a normal (Gaussian) distribution, the use of anormal (not normal) distributions will reduce the efficiency of the parametric tests and likely result in the acceptance or rejection of the wrong hypothesis (Webster, 1977). Some soil properties such as soil potential (Webster, 1966) and hydraulic conductivity (Nielsen *et al.*, 1973) are significantly skewed and consequently the geometric mean is a better approximation of the 'typical' sample. When anormal distributions cannot be 'normalized' by a transformation (i.e. logarithmic, square root, etc.), nonparametric methods, which do not make stringent assumptions regarding normal distributions, should be used (McIntyre and Tanner, 1959; Heath, 1979). Normal and log-normal distributions appear to be the most frequently observed statistical distributions for describing the spatial variabilty of soil physical properties (Rao *et al.*, 1979).

Numerous statistical tests are available for testing the 'normality' of the distribution and should be utilized not only to provide a description of the population dispersion, but to also increase the accuracy of statistical comparisons. Rao *et al.* (1979) found that the two most frequently used methods for establishing normality or log-normality have been (i) visual inspection for skewness using an histogram (Webster, 1966; Cassell and Bauer, 1975; Nielsen *et al.*, 1973) and (ii) examination of the fractile diagram obtained by plotting the measured values on probability paper (McIntyre and Tanner, 1959; Rogowski, 1972; Biggar and Nielsen, 1976; Gumma'a, 1978). The main disadvantage to these methods is that they are not based upon quantitative measures and therefore an objective evaluation of the goodness-of-fit of the theoretical distribution to the measured data is not possible (Rao *et al.*, 1979).

Skewness and Kurtosis

The variance, the second moment about the mean (m_2), of a set of observations describes the amount of spread of the population. The third moment (m_3) moment is the skewness which describes the symmetry of the distribution. The fourth moment (m_4) is kurtosis which describes the peakedness (Webster, 1977).

The third moment may be computed according to equation 17.

$$m_3 = \frac{1}{N} \sum_{i=1}^{n} (x_i - \bar{x})^3 \qquad (17)$$

Where N is the total number of observations; x_i is an individual observation; and $\bar{x}$ is the mean of the observations.

The expression of skewness as a dimensionless quantity is referred to as the *coefficient of skewness* and is denoted as $\sqrt{b_1}$ or λ_1, which may be calculated by:

$$\sqrt{b_1} = \lambda_1 = \frac{m_3}{m_2 m_2^{1/2}} \qquad (18)$$

The fourth moment about the mean, or kurtosis, is:

$$m_4 = \frac{1}{N} \sum_{i=1}^{n} (x_i - \bar{x})^4 \qquad (19)$$

from which the dimensionless quantity b_2 may be calculated.

$$b_2 = m_4 m_2^{-2} \qquad (20)$$

For a normal distribution this ratio is equal to 3. The quantity λ_2 is defined equal to b_2 - 3 giving $\lambda_2 = 0$ for a normal distribution. Distributions more peaked than normal distributions have positive values of λ_2 and distributions flatter than normal have negative values. For further discussion of these parameters refer to Webster (1977) and Snedecor and Cochran (1980).

A test for departure from normality based upon the $\sqrt{b_1}$ and b_2 parameters is given by Bowman and Shenton (1975). The calculated parameters are used to estimate the probability that the data satisfies a normal frequency distribution using isopleth probability figures.

Another test for normality is the W statistic test by Shapiro and Wilk (1965). This test uses an analysis of variance to compare the squared slope of the probability plot regression line of the ordered observations against the expected values from an hypothesized normal distribution. The following summary of the computational methodology is from Shapiro and Wilk (1965).

The object of the W test is to provide an index or test statistic to evaluate the normality of a set of observations. The statistic is an effective measure of normality even for small samples ($n<20$).

The following summary of the computational methodology is from Shapiro and Wilk (1965).

1. Order the observations from small to large.
2. Compute the sum of squares, S^2.
3. Subtract the smallest observation from the largest and multiply the result by a coefficient (provided in Table 5 in Shapiro and Wilk, 1965). Repeat for the second smallest and second largest. Repeat this procedure until all differences have been found and then sum the results to calculate b.
4. Compute $W = b^2/S^2$.

Small values of W are significant, i.e. indicate non-normality. Tabulated values of W are provided for 1, 2, 5, 10, 50, 90, 95, 99% points of the distribution. For example a calculated W of 0.986 for $n=42$ would fall between the 90 and 95% levels indicating that there is at least a

90% probability that the sampled population is normal.

4.2.2 REGIONALIZED VARIABLES

4.2.2.1 Theoretical

A continuously distributed property displays continuity from point to point in space. The value measured at one point is related to the values at adjacent points. The greater the distance between the sample points, the weaker the relationship. If a variable changes from one point to another with apparent continuity, but in a manner too complex to be represented by an ordinary workable function, it is termed a *regionalized variable* (Davis, 1973). The theory of regionalized variables was developed by a French geostatician, Matheron (1965; cited in Journel and Huijbregts, 1978). Several other geostatistical researchers have developed it further with practical applications (Blais and Carlier, 1968; Olea, 1977; Journel and Huijbregts, 1978). It has been extensively applied in soil science by Burgess and Webster (1980a,b), Vieira *et al*. (1981, 1983) and others.

A regionalized variable has the following characteristics (Olea, 1977):

1. an observation which is a value from a function whose argument contains geographical coordinates specifying the location where the observation was made;
2. an average continuity in a mathematical sense. The spatial variation can be great or small but continuity

must exist from point to point; and

3. a random or stochastic component which has no continuity.

A regionalized variable may show different kinds of anisotropy in which variation may not be as great along one direction as along another (Matheron, 1965; cited in Journel and Huijbregts, 1978).

The main assumption used with regionalized variables is that of "stationarity". A series of measurements is stationary or homogeneous if measurements from any of its parts are representative for the entire domain (Agterberg, 1974). Thus all the random variables within the domain have the same mean and variance. If the mean changes with distance, it is referred to as non-stationary.

Weak stationarity occurs when all variables of a set of observations (series) have the same mean, variance, and autocorrelation function. In most practical applications it is sufficient to assume weak stationarity or *second order* stationarity. The series is considered "strictly stationary" if all higher-order moments remain equal (Agterberg, 1974; Journel and Huijbregts, 1978).

A method used to describe the degree of similarity between points separated by a measured distance is the semivariogram. This method uses an approach similar to moving averages to represent changes in variation with changes in distance.

4.2.2.2 Semivariograms

The semivariance describes the average rate of change of variation over distance by calculating the variance for all samples the same distance apart. By definition the semivariance is:

$$\gamma(x,d) = E\{[Z(x_i) - Z(x_i+d)]^2\} \qquad (21)$$

which in turn can be estimated by

$$\gamma^*(d) = \frac{1}{2} N(d) \sum_{i=1}^{N} [z(i)-z(i+d)]^2 \qquad (22)$$

in which E is the expected value; x_i identifies a coordinate position either in space or time; and $N(d)$ is the number of pairs of observations $[z(i), z(i+d)]$ separated by a distance or lag vector d (Journel and Huijbregts, 1978).

The semivariance, $\gamma^*(x,d)$, is a function of both the point x and the vector d. The estimated semivariance is the arithmetic mean of the squared differences between the two experimental measures; $[z(i), z(i+d)]$. A plot of $\gamma(d)$ versus the corresponding values of d is called a semivariogram. By definition $\gamma(d) = 0$, when d, = 0, and as d increases the semivariogram increases to a maximum and, ideally, maintains this level at larger distances (Figure 19).

The semivariogram of a spatially dependent variable (such as that displayed in Figure 19) provides several values which can be of use in quantification of the dependence. The *sill value* (C) is the value of $\gamma(d)$ which remains constant with increasing d; the *zone of influence*

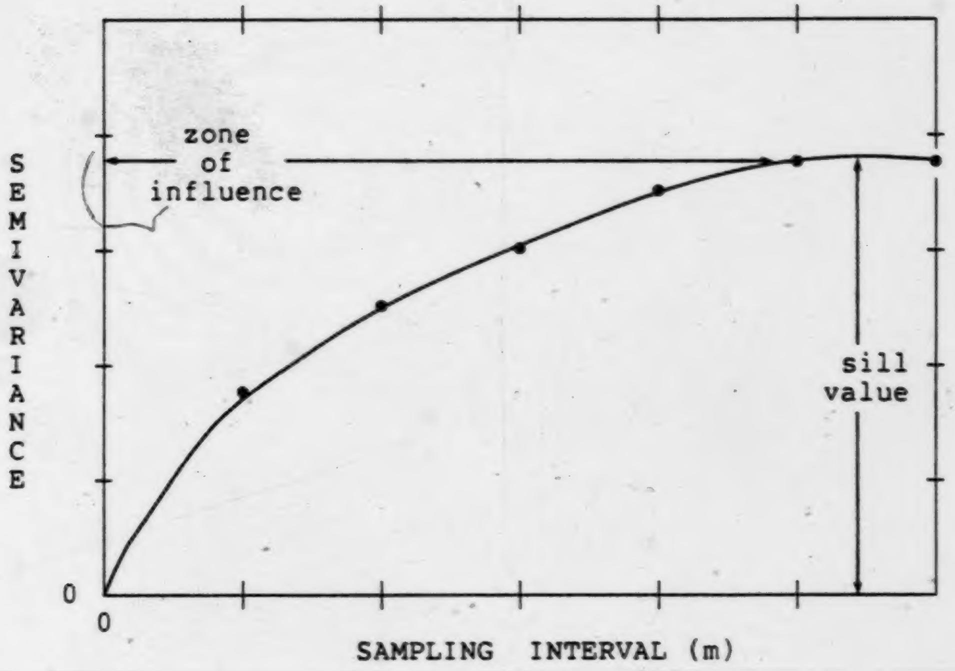

FIGURE 19. IDEAL (SPHERICAL) SEMIVARIOGRAM

(*a*), at which the value *C* occurs, is the distance at which samples become independent of one another; and when $\gamma*(d)$ does not approach zero as *d* approaches zero, the corresponding $\gamma*(d)$ value is the *nugget value* (*Co*) (Matheron, 1963). The *C* value offers an estimate of minimum distance for spacing of independent samples (Campbell, 1978). If a *Co* value greater than zero occurs, then the sample interval is larger than the zone of influence and all variation is not accounted for.

In practice $\gamma*(d)$ is rarely calculated past about half the total distance sampled (Clark, 1979).

Generally there are four characteristic shapes of semivariograms (Davis, 1973):

a) Spherical or exponential (Figure 20a). This type is characteristic of a regionalized variable with high continuity and of which the zone of influence is within the sampling distance.

b) Linear (Figure 20b). This type is characterized by a linear relationship between $\gamma*(d)$ and *d*, and is typical of variables with weak continuity. The lack of a sill indicates that perhaps the total sampling distance is too small to define the zone of influence.

c) Discontinuity at origin (Figure 20c). $\gamma(d)$ does not tend to zero and represents a variable that exhibits a "nugget effect". The nugget effect is due to both discontinuous variables and to microvariabilities as the distance between sample points is too great to measure

SEMIVARIOGRAMS

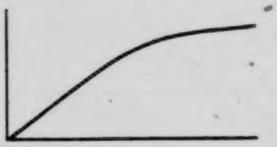

FIGURE 20a
S1. Spherical, high continuity

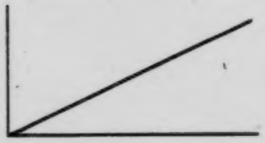

FIGURE 20b
S2. Linear, low continuity

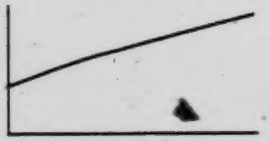

FIGURE 20c
S3. Discontinuity at origin

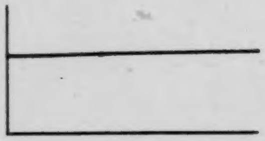

FIGURE 20d
S4. Pure nugget effect

AUTOCORRELOGRAMS

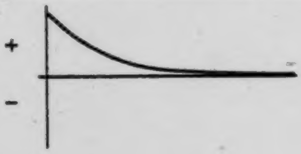

FIGURE 20e
A1. High continuity

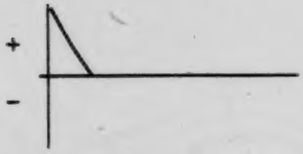

FIGURE 20f
A2. Discontinuity at origin

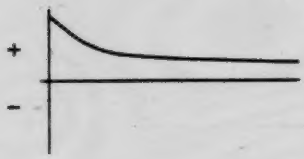

FIGURE 20g
A3. Low continuity

FIGURES 20a to 20g. Common semivariograms and autocorrelograms the total variability. Discontinuity at zero may occur for both of the previous types of semivariograms.

d) Pure nugget effect (Figure 20d). This type corresponds to a totally random variable. The sample interval, *d*, is larger than the zone of influence.

A mathematical function fitted to the line to describe the relationship of variance and distance may be used to statistically interpolate other points through "kriging". "Kriging is the process of estimating the value of a spatially distributed variable from adjacent values while considering the interdependence expressed in the variogram." [1] With this method a more efficient estimate of the mean and variance may be obtained with fewer samples (Webster and Burgess, 1984) and the property may be more accurately mapped than with other interpolation techniques (Vieria *et al*., 1981).

4.2.2.3 Autocorrelation

Another method of graphically depicting spatial dependence is with the use of autocorrelograms. Basically this method measures how the correlation between pairs of terms, *d* units apart, varies with distance. Given *n* values, x_1, x_2, ..., x_n, the so-called serial correlation of lag *d* is calculated by:

---

[1] Davis, 1973, p 385.

$$r_d = \frac{\frac{1}{n-k} \sum_{i=1}^{n-d} (x_i - \bar{x})(x_{i+d} - \bar{x})}{\frac{1}{n} \sum_{i=1}^{n} (x_i - \bar{x})^2} \qquad (23)$$

This form is used only for series of moderate length (at least 50) while a longer form is used for short series for exact estimates and to avoid values of r_d greater than unity (Davis, 1973).

The array of coefficients r_0, r_1, r_2, ... , tells the nature of the internal dependence of the series. Their totality is called the correlogram (Davis, 1973). Typical correlograms are displayed in Figure 20.

4.3 LITERATURE REVIEW

4.3.1 SOIL PROPERTY VARIATION

4.3.1.1 Introduction

The amount of variability that occurs in any measurement is due to several sources: location, measurement, and temporal (Cameron *et al.*, 1971). Although temporal conditions can cause changes in soil moisture, the main source of variability remains that of locational. Measurement variability is generally viewed as minor (Reed and Rigney, 1947; Hewlett *et al.*, 1964; Cameron *et al.*, 1971; Sinclair and Williams, 1979).

4.3.1.2 General Variability of Soil Water Properties

Typical ranges of soil property CV's are presented in

Table 10. The data within this table have been obtained from a variety of sources and grouped with no attempt to separate sources of variability due to different soils, vegetation, climate or sizes of sample area. The groupings of low, medium, high, and extreme correspond with the typical range of CV's encountered for these properties. For example high CV's of 20 to 30% for moisture content have been reported (Towner, 1968; Bell *et al.*, 1980; Hawley *et al.*, 1983) but these are due to large sampling areas encompassing several soil map units or cracking clays. Low CV's of 20 to 30% for unsaturated hydraulic conductivity were reported by Stockton and Warrick (1971) for a 40 ha field from 36 core samples. This could be due to their use of moisture release curves, with a CV of 16% at 10 kPa, with the Millington-Quirk method (Millington and Quirk, 1959) to calculate hydraulic conductivities.

Changes in soil moisture content have relatively high CV values. CV's of 24 to 39% for changes in soil moisture content due to drainage from 4 neutron monitoring sites within a 50 m^2 area were reported by van Bavel and Stirk (1967). Nielsen *et al.* (1973) reported CV's of 69 to 92% for drainage fluxes immediately after cessation of ponding for 20 sites scattered on a 150 ha field. These values dropped to 20 to 50% after several days when the drainage rates slowed. McGowan and Williams (1980a) obtained CV's of 60 to over 200% for changes in soil moisture due to evapotranspiration and drainage from eight closely spaced access tubes in a field of barley. Sinclair Williams (1979) reported CV values for changes in soil moisture content, as measured by the neutron probe, ranging from 6.3 to 16.7% for three fields ranging in size from 0.1 to 0.5 ha.

The general grouping of these soil properties in Table 8 is also confirmed by similar groupings by Beckett and Webster (1971), Warrick and Nielsen (1980), and Wilding and Drees (1983).

The type of parent material affects the relative degree of variability. Mausbach *et al*. (1980; cited in Wilding and Drees, 1983), in a variability study of morphologically matched pedons, observed the following generalized order of spatial variability:

loess < glacial drift < alluvium ≃ residium

Drees and Wilding (1973; cited in Wilding and Drees, 1983) suggested the following generalized array of spatial variability for physical, chemical and elemental properties of parent material:

Loess < glacial till < glacial outwash ≃
glacial lacustrine ≃ alluvium

Several trends in soil variability have been noted:

1. variability increases with depth (Cameron *et al* ., 1971; Beckett and Webster, 1971; Wilding and Drees, 1983);
2. variability increases with distance (Petersen and

TABLE 10. Summary of coefficients of variation for select soil water properties

PROPERTY	RANGE	MEAN	COMMENTS
LOW VARIABILITY			
Bulk density	3 - 17	8	
Moisture content	4 - 42	12	
MEDIUM VARIABILITY			
Field capacity	4 - 30	16	cores
	4 - 52	20	pressure plate
1500 kPa	8 - 54	25	
Available water	9 - 56	20	
Sand	2 - 73	31	
Silt	10 - 79	32	
Clay	8 - 53	32	
HIGH VARIABILITY			
Tensiometers	-	69	Webster (1977)
Infiltration rates	12 - 130	62	
K saturated	24 - 561	94	
K unsaturated	106 - 459	272	Nielsen *et al.*, 1979
Diffusion		6.5×10^6	Hillel, 1980b
Pore water velocity		1.1×10^3	Hillel, 1980b

Where not indicated, coefficient of variation values are obtained from numerous literature sources.

Calvin, 1965; Beckett and Webster, 1971; Wilding and Drees, 1983); and

3. variabilty changes with time due to the effects of climate, vegetation and other factors (Cameron *et al.*, 1971; Gifford, 1979; Hawley *et al.*, 1983).

4.3.1.3 Effect of Depth

Beckett and Webster (1971) reported slight increases in variability with depth for most properties. Cameron *et al.* (1971) confirmed this for soil nutrients but postulated that this could be due to small quantities near the resolution of the instrument. Coelho (1974) found that the variability of silt and clay content, and the amount of water retained in soil samples at specific pressures, increased with depth. Gumma'a (1978) reported increases in variabilty with depth for soil water retentions but texture had only slight increases in CV with depth. Increases in variability with depth has been also reported for moisture retentions by Nielsen *et al.* (1973); Parkes and Waters (1980); Cassel and Bauer (1975), for moisture contents by Nielsen *et al.* (1973); Guma'a (1978), for texture by Mader (1963); Webster (1975), and for hydraulic conductivity by Guma (1978); Russo and Bresler (1981). Exceptions, however, occur.

Towner (1968) reported that the variability of moisture content and available water capacity decreased with depth. This was due to extreme cracking of the surface soils from drying. Moisture contents have been reported to decrease in variability in the depths 0 to 15 cm (Bell *et al.*, 1980; Hawley *et al.*, 1983). This could be related to extreme dryness (-1500 kPa) of the immediate surface (0 to 5 cm), coupled with increasing moisture content with depth.

Less weathering of the deeper horizons could be the cause of the greater variabilities. Harradine (1949) found that many soil properties were more variable in "younger" soils than "older", more weathered soils. Schafer (1979) found that mine soils were much more variable than adjacent undisturbed soils. The effects of climate, relief, turbation by biological activity, and chemical and physical weathering likely act as homogenizing influences upon the inherent variability of the parent material. Agricultural activities by man tend to create smaller variations in texture and moisture content. Reynolds (1970b) found that recently cultivated sod and young crops had lower variability in soil moisture than permanent pasture and forest.

4.3.1.4 Effect of Size of Area Sampled

In general the variability of any soil property will increase as the area sampled increases. Table 11 lists field moisture content CV's for areas ranging in size from 0.02 m^2 to over 900 ha.

For very small sample areas (<4 m^2) variation is related to the sample volume. Hawley *et al.* (1983) collected 10 gravimetric samples for each of eight soil volumes (7 to

TABLE 11. SOIL MOISTURE VARIATION AND SIZE OF AREA SAMPLED

CV (%)	AREA	COMMENTS	REFERENCE
VERY SMALL	**0.02 to 4 m sq**		
5.6 - 6.0	.02 - 1	10 samples, depth 5-8 cm, 11% slope	Hills and Reynolds, 1969
8.7 - 15.9	1	sloped land, catchment, England.	Hills and Reynolds, 1969
2.1 - 7.2	2	3 plots, 5 sets of 8-10 gravimetric samples per plot, sample volumes 54-820 cc, cultivated SL, Maryland.	Hawley *et al.*, 1982
SMALL: 4 to 1000 sq m			
12 - 23	4	4 neutron tubes 1 m apart, England	McGowan and Williams 1980a
3.2	50	4 neutron tubes, 420 hours after ponding, 1 m, cultivated CL, Texas.	Van Bavel *et al.*, 1968
4.1 - 8.7	6 - 961	10 samples 5-8 cm depth, 11% slope	Hills and Reynolds, 1969
13.3	36	5x5 grid neutron tubes, 1.1 m apart, mean for 6 periods to 90 cm	Rouse, 1970
10.5 - 23.1	25	cracking black clay soil, 6-8 samples	Towner, 1968
2.7 - 3.0	0.1 ha	15 neutron tubes in cleared site of forest Australia, measured 13 days after flooding	Sinclair and Williams, 1979
5.9	60 m	transect, 1 m interval, 3 days after	Vauclin *et al.*, 1982
9.9	100 m	irrigation, 0-5 cm depth, California	
MEDIUM	**0.1 to 3 ha**		
5.2, 6.1	1, 1.5 ha	8 and 6 gravimetric samples, 15-30 cm, tobacco and corn fields in Florida	Hammond *et al.*, 1956
5.4 - 18.6	0.16 ha	8 neutron tubes, Alfisol, Australia, pasture	Sinclair and Williams, 1979
8.3 - 25.7	0.18 ha	128 neutron tubes, SL, North Conneticut, different tillage treatment, soybean crop	Cassel and Nelson, 1981
17 - 24	1 ha	60 samples, inferred from graph	Hills and Reynolds, 1969
LARGE	**Greater than 3 ha**		
10 - 21	5.3-17.9	agricultural watersheds, Olkahoma, 0-15 cm	Hawley *et al.*, 1983
27 - 35	11	20 samples, 5-8 cm, 'drainage basins	Hills and Reynolds, 1969
23 - 29	100	20 samples, inferred from graph	Hills and Reynolds, 1969
20 - 28	900	20 samples, inferred form graph	Hills and Reynolds, 1969
3 - 11	4- 17	sandy agricultural soils, 12-24 gravimetric samples, 15-30 cm, Florida	Hammond *et al.*, 1956
6 - 24	16	71 fields in Kansas, Arizona, and N. Dakota 19-36 gravimetric per field, 0-10 cm	Bell *et al.*, 1980
26 - 50	16	3 fields, 56 sites per field, gravimetric samples all depths, 3 days after irrigation	Gumaa, 1976

824 cm^3) from two m^2 plots. For volumes greater than 50 cm^3, no further decrease in variation occurred. Smaller volumes showed relatively high variation. Sisson and Wierenga (1981) studied infiltration rates of five 5 cm diameter rings nested within each of five 25 cm diameter rings nested in turn within each of thirty-six 127 cm rings arranged in a six by six m grid. Approximate CV's obtained along 1 large ring transect were: five 127 cm rings - 43%; twenty five 25 cm rings - 82%; and one hundred and twenty five 5 cm rings - 91%. Beckett and Webster (1971), from an extensive literature review, found that up to half of the variability for most soil properties that occurs in a field may be found in any 0.5 m^2 of it.

Variation on a small scale defined here refers to plot sizes ranging from 4 to 1000 m^2 that are usually located on uniform soils and topography. CV values of 12 to 23% for 4 neutron tubes 1 m apart (McGowan and Williams (1980a) are higher than other reported values. Unfortunately no background or explanation was provided.

Soil hydraulic properties exhibit large variability over short distances. Babalola (1978) observed that CV's for hydraulic conductivity were only slightly smaller for a 0.3 ha plot than a 92 ha field. Byers and Stephens (1983) obtained CV's for saturated hydraulic conductivity of 39 and 62% for 92 samples from each of two transects 14.8 m long. A CV of 130% was reported for 48 infiltration rings arranged in a 6 by 6 m plot on subsoil material (Luxmoore *et al.*, 1981).

Hills and Reynolds (1969) in a study of soil moisture variability in the upper 5-8 cm, for plot sizes ranging from 1 m^2 to a catchment basin 100 to 1000 ha, found that in terms of variability, two size ranges emerged: small plots up to 961 m^2; and the larger "drainage classes". Reduction of sample area within these bounds did not necessarily result in a reduction of the CV.

Large scale effects upon soil variability are defined as those involving changing soil types, vegetation, parent material, relief, and climate. Reynolds (1970b) found that vegetation affected the amount of variability of moisture content. Cultivated soils had lower variabilities than permanent pastures or forests, which were lower than clumped vegetation. Topography affects soil moisture due to slope, aspect and location on slope. Slope influences infiltration, runoff, water redistribution (Hawley *et al.*, 1983; Hills and Reynolds, 1969).

Wilding and Drees (1983), from a review of variability and soil classification, found that the magnitude of variability generally increased from pedons to polypedons to mapping units of a given series to all soils within the survey area. Most properties (texture, color, depth to carbonates, horizon thickness, pH, organic matter, exchangeable cations, and CEC) in mapping units had CV's between 25 and 40%, while polypedons commonly had values 1/2 to 2/3 of these, and pedons exhibited CV's of 5 to 10%.

4.3.1.5 Temporal effects upon variation

Soil moisture content changes with time due to drainage, evapotranspiration, and precipitation. Some studies reported that moisture content variability increases with dryness (Towner, 1968; Guma'a, 1978; Sinclair and Williams, 1979); others that it decreased with dryness (Reynolds, 1970a; Nielsen *et al.*, 1973; Bell *et al.*, 1980; Hawley *et al.*, 1982). McGowan and Williams (1980a) list three reasons why CV's can be high for moisture contents:

1. nonuniform wetting due either to hydrophobic surfaces or soil cracks. Measured changes following a storm are often highly variable;
2. point variation in evaporation rates are particularly noticeable where soils are approaching their maximum soil water deficit, especially where germination is poor; and
3. variation in drainage can occur especially where there are abrupt textural changes or recession in water tables.

Hawley *et al.* (1983) suggested that after a heavy rainfall most of the water is initially taken up in the larger pores which usually constitute a small percentage of the total pore volume. These larger pores, due to structural properties, and possibly turbation by meso- and macro-fauna account for the greater variability of moisture content. With time the soil moisture in these large pores equilibrates with that in the smaller pores and the soil moisture variability reduces.

The increase in CV reported by Towner (1968) as the soil dried was due to cracking. Varazashika *et al.* (1976; cited by Guma'a, 1978) indicated that the CV increased markedly for decreasing moisture content and that this was due to the role of clay surface area at potentials of approximately -1500 kPa. At those potentials moisture content is related more to surface area of the clays, whereas at higher potentials moisture content is held by capillary forces. Webster (1966) reported that the CV was lowest in freely drained soils and higher in soils with a seasonably high water table or with a fine texture.

4.3.2 SPATIAL DEPENDENCE OF SOIL PROPERTIES

4.3.2.1 Introduction

Application of regionalized variable theory in soil science has served two main purposes; to aid in the delineation of soil map boundaries (Webster and Cuanalo, 1975; Lanyon and Hall, 1981; McBratney and Webster, 1981); and to obtain a better understanding of the causes of soil variability in small study sites (Gajem *et al.*, 1981; Vieria *et al.*, 1981). Of particular concern to many researchers and to the objectives of this study is the description and understanding of variability of soil water properties that occurs within a small area (less than 1 ha) (Luxmoore *et al.*, 1981; Vieria *et al.*, 1981; Byers and Stephens, 1983; McBratney and Webster, 1983). As discussed in the previous section, the variability of some of the soil water properties can be quite high, even within small plots with the same soil type, relief, vegetation and management practice. The effort of many researchers has been directed towards obtaining a mathematical description of the spatial structure and an explanation of the causes of the variability in an effort to obtain more precise estimates of desired parameters with fewer samples.

4.3.2.2 Scale and Spatial Dependence

The design of a proper sample plan must fulfil two sampling criteria; how large an area needs to be sampled; and what the intensity and the spacing of samples should be. Burrough (1981, 1983a, 1983b) considered the separation of systematic variation and random variation ('noise') to be entirely scale dependent because altering the scale of observation almost always revealed systematic structure in the noise. He stated that the "white noise concept of a normally distributed random function must be replaced to take into account the nested, autocorrelated and scale-dependent nature of unresolved variations" [2].

The significance of nested scales is revealed in the semivariances and autocorrelation data for certain soil properties presented in Table 12. Gajem *et al.* (1981) found different sill values for -10 and -1500 kPa moisture release values for three different sampling intervals (*d*): 20, 200,

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[2] Burrough, 1983b, p 582.

TABLE 12. SPATIAL DEPENDENCE FOR SELECTED SOIL PROPERTIES

Property	*Co* (m)	**Model**	*d* (m)	*L* (m)	**Comments**	**Ref.**
Bulk density	3.4	A1	0.2	20	depth 50 cm	1
	2.0	A2	2.0	200	typic torrifluevent	1
Moisture	2.4	A1	0.2	20		1
	>46	A3	2.0	200		1
	40	S3	1.0	100	depth 0-5 cm, irrigated	2
	6	A1			east-west transect	
	3	A1			north-south transect	
	6	A2	3.0		6x6 neutron probe grid deciduous forest.	3
	<50	A2	50.0		8x8 gravimetric grid. 2 days after irrigation	4
	<9	S4	9.0		approximate grid in rangeland watershed, 0-15	5
	39,23,25,21	A1	random		depths; 0 30 60 90 cm; 30 random samples in 120x60m area, Rhodoxeralf	6
Moisture, saturated	76,73,45,28	A1	random		as above	6
Moisture, 10 kPa	3.5	A1	0.2	20	one of three transects in 85 ha field	1
	1.0	A2	0.2	20		
	0.6	A2	0.2	20		
	40,>46,>40	A3	2.0	200	as above 85 ha field	1
	160	A1	20.0	2000		1
1500 kPa	1.7	A1	0.2	20	as described above	1
	0	A2	0.2	20		
	40,>40	A3	2.0	200		
	36,14	A1	2.0	200		
	150	A1	20.0	2000		
	40	S2	10.0		7x4 grid	7
Water table	>2 km	S2	250.0		grid, 2.25x3 km	8
Texture						
C, Si, S	>5	A3	0.2	20		1
C	<50	A2	50.0		8x8 grid	4
C	36	S1	10.0		7x4 grid	7
Si	50	S2				7
S	36	S1				7
S	30	-	10.0		grid 80x200 m Aquic argiudoll	9
S	<40	-	10.0		same grid but on Pachic Arguistolls	9
Si	>100	S2a	10.0		topsoil grid, 10.6 ha	10
Si	>100	S2b	10.0		subsoil	
Infiltration	50	S1	1.0		grid 160x80 m, with rows 1,5,15, and 19 m Typic Xeroorthents	11
	<2	S4	2.0		6x8 m grid, subsoil Typic Hapludult	12

Model type; *A* refers to autocorrelogram and *S* to semivariogram. Refer to text in Statistical Methods for explanation of model type.

a refers to distance of spatial dependence; *d* is sample interval

L is length of transect

References:

1-Gajem *et al.*, 1981; **2**-Vauclin *et al.*, 1982; **3**-Clarke, 1976; **4**-Gumaa, 1978; **5**-Hawley *et al.*, 1983; **6**-Russo and Bresler, 1981b; **7**-Vauclin *et al.*, 1983; **8**-Dahiya, 1979; **9**-Campbell, 1978 **10**-McBratney and Webster, 1983; **11**-Vieria *et al.*, 1981; **12**-Luxmoore *et al.*, 1981.

and 2000 cm. McBratney and Webster (1981) found a change in slope of the semivariogram for soil color, pH, and depth of topsoil at intervals of 8 and 160 m, suggesting a change in variation of scale. Byers and Stephens (1983), from saturated hydraulic conductivity rates obtained every 15 cm along two 14.85 m transects, found sill values at 0.15 and 0.60 m for natural log values of the hydraulic conductivities. Russo and Bresler (1983), from 30 random samples in an 80 by 120 m field, obtained spatial dependence for hydraulic conductivity ranging from 14 to 39 m dependent upon the depth. Reynolds (1970ac) noted that variability of soil moisture appeared to be constant for the small plot size (under 450 m²) and changed for 'drainage classes' indicating several levels of variability. Soil moisture variability was reported by Gajem *et al.*. (1981) as having two ranges of dependence, 2.4 m and greater than 46 m. Large nugget values for the infiltration rates reported by Luxmoore *et al.* (1981) and for moisture contents reported by Hawley *et al.* (1983) indicated that samples should be located closer together in order to discern structure.

4.3.2.3 Anisotropy in Spatial Variability

Anisotropy in the variability of soil properties have been reported by Byers and Stephens (1983) for hydraulic conductivity and by Vauclin *et al.* (1983) for soil moisture and soil temperature. The anisotropic effect on soil moisture and temperature were attributed to wind effects during irrigation.

4.3.2.4 Sampling Precision

Several studies have used regionalized variables to obtain better estimates of variability and to reduce sampling intensity. Vieira *et al.* (1981) measured infiltration rates at 1280 locations in a field 160 by 55 m. Using the spatial dependence revealed by semivariograms, they concluded that 128 samples would be enough to obtain the same information. McBratney and Webster (1983a) and Vauclin *et al.* (1983) used cross semivariograms to express the spatial relations among two interdependent properties. The precision of a less intensely sampled or more variable property can be improved through co-kriging it with a more precise interdependent property. McBratney and Webster (1983a) found strong co-regionalization to exist between topsoil silt, and subsoil silt and subsoil sand. This enabled topsoil silt to be estimated more precisely by co-kriging than by kriging from data on topsoil silt alone. They suggest that when the auto and cross semi-variograms for a set of variables are known in advance they can be used to plan an optimal sampling plan by sampling the main variable on a larger grid than that for subsidiary variables.

McBratney and Webster (1983b), using known semivariograms for several soil properties coupled with kriging, reduced sample sizes required for a desired precision to 11 to 28% (i.e. increased efficiency) of that calculated by classical theory for simple random sampling.

Webster and Burgess (1984) discussed improved sampling schemes for small parcels of land using the semivariogram. Depending upon the shape of the semivariogram, the actual sampling design within each block will change to maximize precision. Composite sampling was shown to increase precision of the estimate of the semivariogram.

Regionalized variable theory has been shown to increase precision and decrease sampling effort, provided that a spatial structure exists within the scale sampled. Even without the use of these geostatistical techniques, sampling on a grid will yield a more accurate estimate of the true mean with a greater precision than that achieved by simple random sampling for most soil conditions (Webster and Burgess, 1984).

4.4 MATERIALS AND METHODS

4.4.1 SAMPLE SIZE ESTIMATION

The following equation was used to estimate sample size for soil properties measured in this study:

$$n = \frac{t_k^2 s^2}{D^2} \qquad (24)$$

Only one level of probability, k, for the t distribution was chosen; 0.95 for two-tailed tests. Two levels of precision, D, were used: 5 and 10% of the mean. The variance, S^2, value used was that estimated from the actual measurements.

4.4.2 TESTS OF NORMALITY

Skewness ($\sqrt{b_1}$) and kurtosis (b_2) parameters were calculated for most soil properties in this study. They were tested for departure from normality using isopleth probability figures by Bowman and Shenton (1975). The calculated parameters were used to estimate the probability that the data satisfy a normal frequency distribution. The isopleth figures present the 90, 95, and 99% probabilities that the tested data are anormal. The contours on the figures are for samples sizes ranging from 20 to 1000. For the purposes of testing sample sizes in the isopleth, values of 20 and 40 were used for actual sample sizes of 21 and 42. The data was tested for anormality at a probability of ≥95%.

The W statistic (Shapiro and Wilk, 1965) was calculated and compared against tabulated values at the 1, 2, 5, 10, 50, 90, 95, and 99% probabilities of "non-normality". The data were tested for non-normality at the probability levels of ≤5% and ≥95%.

4.4.3 DETERMINATION OF SPATIAL DEPENDENCE

The semivariogram was used for the determination of spatial dependence other than the autocorrelogram. The semivariogram offers the advantages of smaller sample size requirements, less stringent assumptions regarding stationarity (more robust), and it provides the basis for statistical spatial interpolation.

Semivariances were calculated for four vectors across the grid: north-south, east-west, northeast-southwest, and northwest-southeast. All semivariances, $\gamma*(d)$, for a specific sample interval, d, for a specific vector were summed and averaged. If anisotropic conditions were not obvious (one vector at least twice the other vector) the vectors with similar d were summed and averaged. The resulting $\gamma*(d)$ values were used to plot a semivariogram (Table 13).

For bulk density, particle size analysis, moisture contents at -33 and -1500 kPa, semivariances were calculated for the entire site for 42 sample locations. For moisture contents, and change in moisture stored, semivariances for the entire site were only calculated before any significant differences in moisture occurred from the barley growth. After barley establishment, semivariances were calculated for only half the site; the three fallow plots alternating with the three barley plots. Semivariances were for these conditions were calculated only for the N-S vectors.

TABLE 13. Site semivariogram vectors and sampling intervals

VECTOR	PARAMETER						
N-S	*d* (m)	6.1	12.2	18.3	24.4	30.5	36.6
	n	36	30	24	18	12	6
E-W	*d* (m)	6.1	12.2	18.3	24.4	30.5	
	n	35	28	21	14	7	
NE-SW	*d* (m)	8.6	17.3	25.9	34.5	43.1	
	n	30	20	12	6	2	
NW-SE	*d* (m)	8.6	17.3	25.9	34.5	43.1	
	n	30	20	12	6	2	

d sample interval for calculation of semivariance
n number of sample intervals at specific *d*

4.5 RESULTS AND DISCUSSION

4.5.1 PHYSICAL PROPERTIES

4.5.1.1 Precision

The mean, CV, and the required number of samples for a desired precision for the soil properties; bulk density, texture, and moisture contents at -33 and -1500 kPa are listed in Table 14 by depth. The CV for bulk density was the highest (9.3%) at the 40 to 60 cm depth, whereas sand, silt and clay, had the highest CVs' at the 60-100 cm depths.

Bulk density was the least variable, requiring only one to three samples to achieve precision within ±10%, whereas clay and moisture content at -33 kPa for the 80 to 100 cm depth had the greatest variability, requiring 16 samples to achieve a precision within ±10%. If the precision is halved to ±5%, the number of samples increases to approximately 64.

There is a slight tendency for the variability to increase with depth from 0 to 60 cm for bulk density and moisture content at -33 kPa, and from 20 to 100 cm for sand, silt, and clay. A slight decrease in the variability for texture occurred from the 0-20 cm to the 20-40 cm depths.

The depth of the Ah horizon showed a high variability with a CV of 35.3% and sample sizes for desired precisions at 5 and 10% of the mean (D05 and D10) of 203 and 51.

4.5.1.2 Distribution

Distribution parameters describing population normality for bulk density, sand, silt, clay, and moisture contents at

TABLE 14. MEAN, COEFFICENT OF VARIATION, AND SAMPLE SIZES FOR SELECTED SOIL PHYSICAL PROPERTIES

DEPTH (cm)	STATISTIC	BULK DENSITY Mg m^{-3}	TEXTURE SAND %	SILT %	CLAY %	MOISTURE CONTENT AT -33 kPa Mg Mg^{-1}	-1500 kPa Mg Mg^{-1}
0-20	Mean	1.1	33.6	39.6	26.1	30.2	18.1
	CV	6.1	11.4	15.1	14.9	6.1	8.9
	D05	8	21	37	36	6	13
	D10	2	5	9	9	2	3
20-40	Mean	1.3	34.8	33.6	30.3	24.9	14.6
	CV	8.7	10.3	14.5	12.8	9.1	10.3
	D05	12	17	21	27	13	17
	D10	3	4	5	7	3	4
40-60	Mean	1.6	39.0	27.0	32.8	22.3	13.8
	CV	9.3	12.4	13.5	14.7	10.0	18.1
	D05	14	27	30	35	16	53
	D10	4	7	7	9	4	13
60-80	Mean	1.7	40.2	27.2	31.6	23.3	13.4
	CV	5.9	19.2	15.6	15.7	11.9	15.0
	D05	6	60	40	41	23	37
	D10	1	15	10	10	6	9
80-100	Mean	1.7	39.6	27.2	32.0	25.6	14.1
	CV	5.9	19.3	14.3	19.9	12.7	16.1
	D05	6	61	34	64	26	63
	D10	1	15	8	16	7	16

D05, D10 = Levels of precision at 5 and 10% of mean
Student's t value (2.021) for 40 degrees of freedom at 0.05 level of significance for two-tailed test.

-33 and -1500 kPa are listed in Table 15. The test for departure from normality by Bowman and Shenton (1976) is slightly more sensitive than the *W* test by Shapiro and Wilk (1965). Most of the anormality occurs below 40 cm, especially for sand, silt, and clay. The moisture contents at -33 and -1500 kPa show the least anormality except at the 60 to 80 cm depth for the moisture content at -33 kPa.

For bulk density at the 40-60 and the 60-80 cm depths, the anormality is the result of a large negative skew. The cause of these skewed values was the variation of the depths the Ah horizon. In two locations the Ah extended to 54 and 94 cm resulting in bulk densities much lower, 1.15 and 1.16 Mg m^{-3}, for the 40 to 60 cm depth, whereas the mean for this depth was 1.60 Mg m^{-3}. The same trend occurred for the 60 to 80 cm depth. This variation in Ah depths was likely due to tree throw as noted by Pawluk and Dudas (1982) in other locations of the local landscape.

The large negative skew for clay content in the 0-20 cm depth is the result of one very low value, 7.9%, whereas the mean clay content for this depth is 26.1%. The significant anormalities for sand, silt, and clay at the deeper depths (below 40 cm) is the result of sand lenses and layers. Generally, however, only a few values produced large skew values. A very low silt value, 9.1%, in the 60-80 cm depth, at one site was responsible for the large negative skew. The mean silt content at this depth was 27.7%. A high sand value, 73.0% (mean 40.2%), at the same location resulted in

TABLE 15. SKEWNESS, KURTOSIS, AND W VALUE OF SELECTED SOIL PHYSICAL PROPERTIES

DEPTH	STATISTIC	BULK	TEXTURE			MOISTURE CONTENTS AT	
(cm)		DENSITY Mg m^{-3}	SAND %	SILT %	CLAY %	-33 kPa Mg Mg^{-1}	-1500 kPa Mg Mg^{-1}
0-20	Mean	1.1	33.6	39.6	26. 1	0.30	0.18
	Skew	0.062	-.3	1.17	-1.8	0.34	.14
	Kurtosis	2.53	2.75	5.60	14.19	2.83	4.58
	normality	/	/	aaa	aaa	/	/
	W	0.964	0.96	0.91	0.76	0.98	0.96
	normality	/	/	aaa	aaa	/	/
20-40	Skew	0.72	0.08	0.05	0.81	-.17	0.12
	Kurtosis	3.177	4.15	2.32	4.67	1.94	2.14
	Normality	/	/	/	a	/	/
	W	0.945	0.94	0.98	0.95	0.95	0.97
	Normality	aaa	/	/	/	/	/
40-60	Skew	-1.78	0.93	0.24	-1.13	0.23	-0.49
	Kurtosis	5.58	5.61	5.63	6.13	2.93	3.51
	Normality	aaa	aa	a	aaa	/	/
	W	0.78	0.95	0.95	0.90	0.98	0.96
	Normality	aaa	/	/	aaa	/	/
60-80	Skew	-2.18	1.76	-1.47	-0.28	-1.48	-0.29
	Kurtosis	10.45	9.21	8.95	3.40	8.44	3.68
	Normality	aaa	aaa	aaa	/	aaa	/
	W	0.83	0.95	0.95	0.90	0.89	0.98
	Normality	aaa	/	/	aaa	aaa	/
80-100	Skew	-0.21	0.40	1.20	-0.75	-0.91	-0.41
	Kurtosis	3.37	5.59	5.83	4.33	4.96	3.37
	Normality	/	aa	aa	a	aa	/
	W	0.98	0.95	0.93	0.95	0.96	0.98
	Normality	n	/	aa	/	/	/

Normality:

a	anormal at 0.90 to 0.95 level of significance
aa	anormal at 0.95 to 0.99 level
aaa	anormal at <0.99 level
n	normal at 0.90 to 0.95 level of significance
/	not significantly normal nor anormal at >0.90

a large positive skew. Even without these anormal values at this site, the texture still varied from a clay to a sandy loam at the 60-80 cm depth.

Despite the anormalities of the texture values, the moisture contents at -33 and -1500 kPa displayed fewer anormal values (Table 15). The large negative skew for the 60-80 cm depth was due to a low moisture value at the same location of the high sand value.

4.5.1.3 Semivariances

Any properties that displayed significant anormality at a significance level of 10% or less are assumed not to meet the requirements of stationarity and consequently semivariances for these conditions are invalid. Although it has been stated that the semivariogram is more 'robust' than the autocorrelogram (Davis, 1973), no quantification or example has yet been found in the literature that defines the limit of 'robustness'. Consequently semivariograms within this study are interpreted carefully and spatial dependence only recognized if it is definitely indicated.

Semivariances for the nugget value (*Co*), and the sill value (*C*), along with the zone of influence (*a*) and the type of semivariogram are listed in Table 16 for bulk density, sand, silt, clay, and the moisture contents at -33 and -1500 kPa. Except for the moisture contents at -33 and -1500 kPa for 0-20 cm, the semivariances displayed large nugget values indicating that there was no spatial dependence at the

TABLE 16. SEMIVARIOGRAM VALUES FOR SELECTED SOIL PHYSICAL PROPERTIES

DEPTH (cm)	VALUE	BULK DENSITY	SAND	SILT	CLAY	FC	PWP
0-20	*Co*	0.010	16.3	*	*	2.6	3.7
	Type	D	D	*	*	C	C
	C	-	-	-	-	5.3	8.3
	a (m)	-	-	-	-	18	24
20-40	*Co*	*	23.6	32.0	*	8.7	5.1
	Type	*	D	D	*	C	D
	C	-	-	-	-	14.5	-
	a (m)	-	-	-	-	24	-
40-60	*Co*	*	*	*	*	9.3	11.9
	Type	*	*	*	*	D	D
	C	-	-	-	-	-	-
	a (m)	-	30	30	30	-	-
60-80	*Co*	*	*	*	*	*	7.7
	Type	*	*	*	*	*	D
	C	-	-	-	-	-	-
	a (m)	-	-	-	-	-	-
80-100	*Co*	0.019	*	*	*	*	12.5
	Type	D	*	*	*	*	D
	C	-	-	-	-	-	-
	a (m)	-	-	-	-	-	-

Symbols: *Co* Semivariance at sample interval (*d*) of 6 m
Type Semivariogram; A-spherical, B-linear, C-linear with large *Co*, D-pure nugget effect.
C Sill value
a Sample interval at which sill value reached.
* Semivariance cannot be properly estimated due to nonstationarity.

intervals sampled.

The semivariances for the moisture contents at -33 and -1500 kPa indicate linear semivariograms; however, this is just for east-west transects. The north-south transects show lack of sill values indicative of spatial independence conditions.

Generally where some spatial dependence existed it was between 18 to 30 m; however, the sill values occurred where there were too few pairs to calculate reliable semivariances and some indicated that the sill was located at distances greater than that of the grid. The large nugget values and the lack of spatial dependence for most of the properties at most depths suggested that the sample interval (6.1m) is too large to properly infer the sill value. Slight differences in semivariances occurred depending upon the vector, north-south versus east-west; however, the differences were not large enough to result in anisotropic conditions.

4.5.2 SOIL MOISTURE

4.5.2.1 Precision

Field moisture contents showed a tendency to increase in variability (as indicated by the sample size for a required precision) for drier conditions that occurred both before the heavy rainfall period (June 16 to July 7) and later during times of high Et rates (Figure 21). During wet conditions the number of samples required for a precision of ±5% was between 4 and 8.

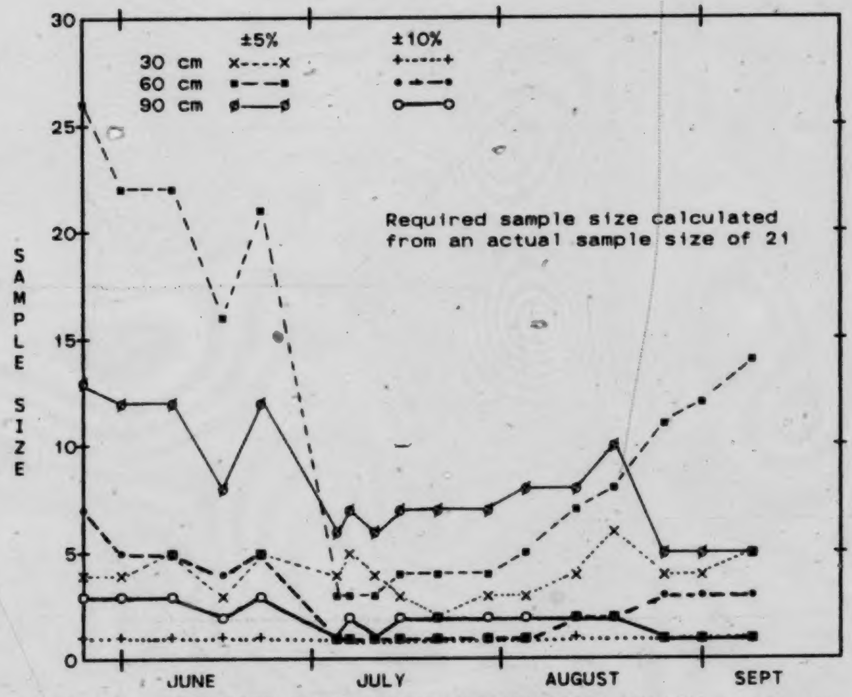

FIGURE 21. REQUIRED SAMPLE SIZES FOR MOISTURE CONTENT AT 30, 60, AND 90 CM IN FALLOW PLOTS FOR PRECISION LEVELS OF ±5% AND ±10%

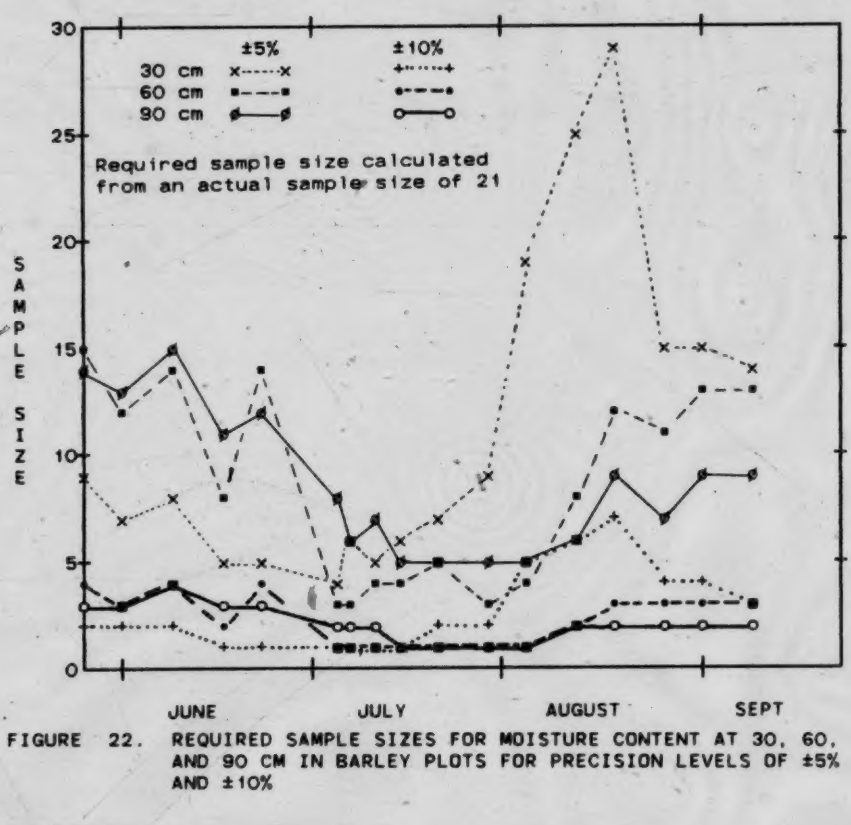

FIGURE 22. REQUIRED SAMPLE SIZES FOR MOISTURE CONTENT AT 30, 60, AND 90 CM IN BARLEY PLOTS FOR PRECISION LEVELS OF ±5% AND ±10%

The moisture content at 60 cm for the fallow plots had the greatest variability before the wet conditions requiring 20 to 25 samples for a precision of ±5%. Generally variability increased with depth for both the fallow and barley plots before June 16. After July 7, during drying due to high Et, the barley plots displayed a reverse trend, with the 30 cm depth having the greatest variability (>25 samples).

For both fallow and barley plots, the number of samples required for a precison of ±10% was consistently below 5 except for the barley 30 cm depth on August 18 when it was 7 samples.

Coefficients of variation for fallow ranged from a high of 12.3% at 60 cm on May 26 to a low of 3.1% for the 15 cm depth on June 16. The 60 cm depth CV was lowest on July 4, at 4.4%, whereas the 15 cm depth CV was highest on August 11, at 8.0%. The highest and lowest CV values for the barley plots were 13.0% at 15 cm on August 11 and 3.3% at 15 cm on June 15. These values are not exceptionally high compared to sites of similar size as reported in the literature (Table 11).

4.5.2.2 Distribution

Distribution of field moisture contents was negatively skewed, especially for the 60 cm depth before the heavy rains during June 16 to July 7 (Figures 23 and 24) . The 30 and 45 cm depths also displayed this trend but to a lesser

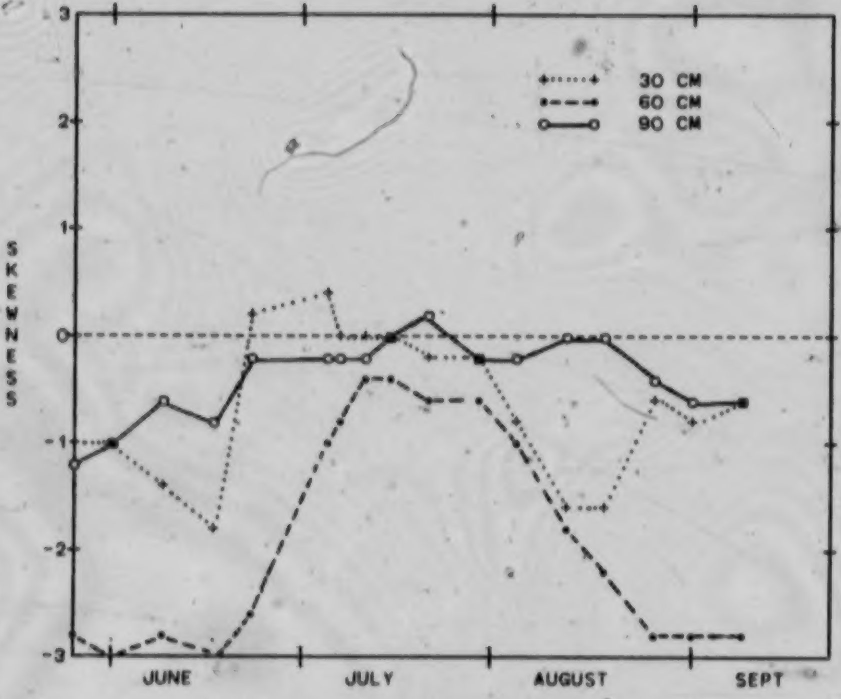

FIGURE 23. MOISTURE CONTENT SKEWNESS FOR FALLOW PLOTS AT 30, 60, AND 90 CM

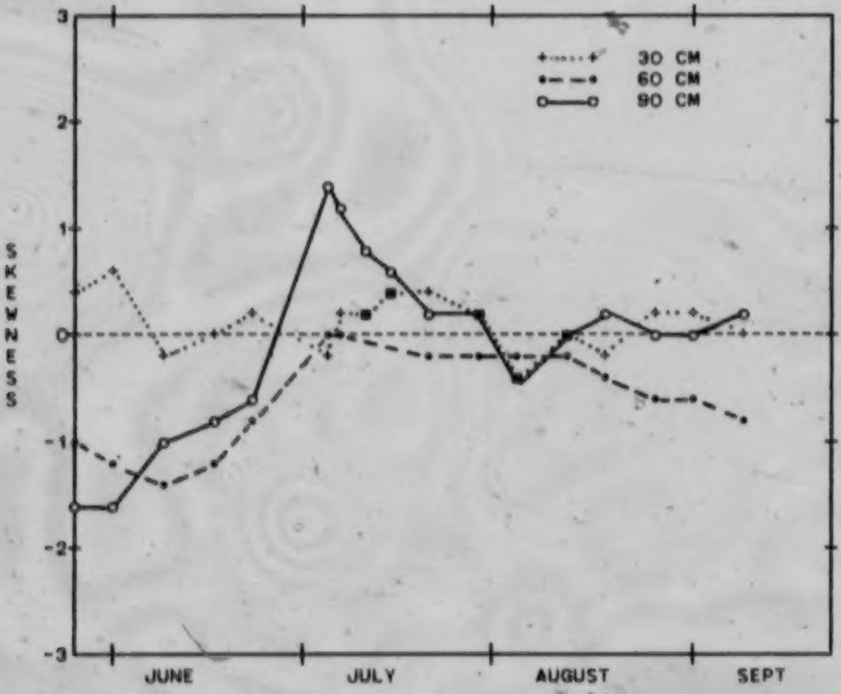

FIGURE 24. MOISTURE CONTENT SKEWNESS FOR BARLEY PLOTS AT 30, 60, AND 90 CM

degree. During the near saturated conditions of late June to early July the skewness tended to zero and slightly positive, except for the barley plots at 90 cm which became significantly positive. During the drying trend of the third period (July to September), the distribution of the moisture content for the fallow plots exhibited a negative skew (Figure 23) and the moisture content distribution in the barley plots remained near 0 skew with the moisture content at 60 cm depth exhibiting a slight negative trend.

The large skew values (-2) for the fallow plots were due to the one site which had a lower moisture content during dry conditions than the average moisture. During wet conditions that site had moisture contents equivalent to the average, resulting in normal distributions. Sand content at this site was anormally high, 73% at 60 cm as compared to the site average of 40%.

W statistic values (Figures 25 and 26) confirm the anormal distributions for fallow dry conditions and the normal distributions during wet conditions.

4.5.2.3 Semivariance

Three dimensional semivariograms for the 30, 60, and 90 cm depths are illustrated in Figure 27 for the fallow plots and in Figure 28 for the barley plots. The plots show semivariance on the vertical (z) axis plotted against the sample interval on the right hand (x) axis and against time on the left hand (y) axis. The semivariograms are plotted

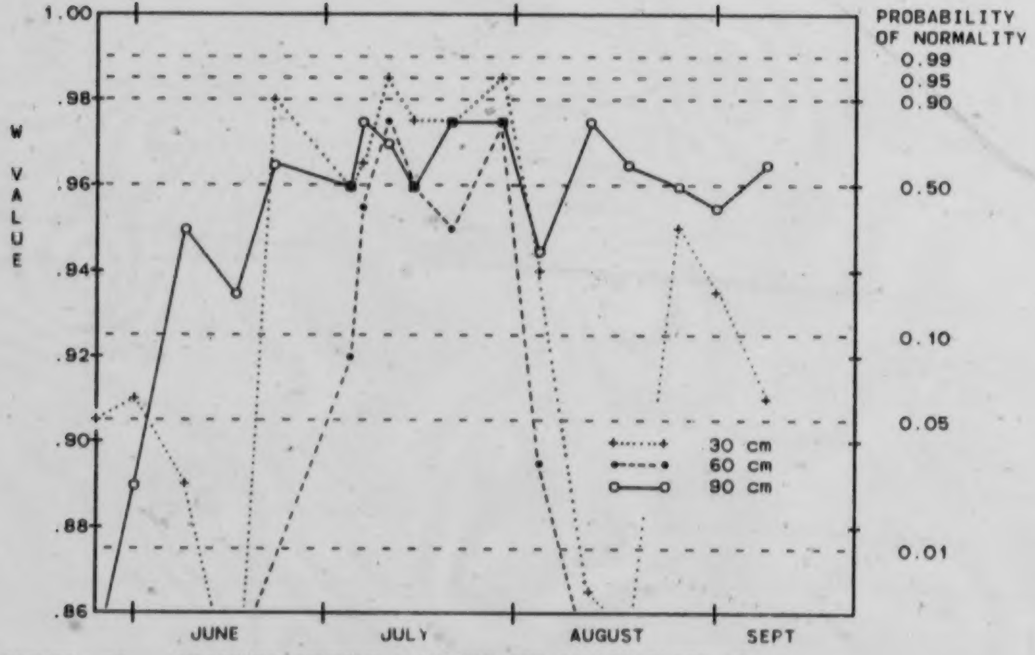

FIGURE 25. MOISTURE CONTENT W VALUES FOR FALLOW PLOTS AT 30, 60, AND 90 CM

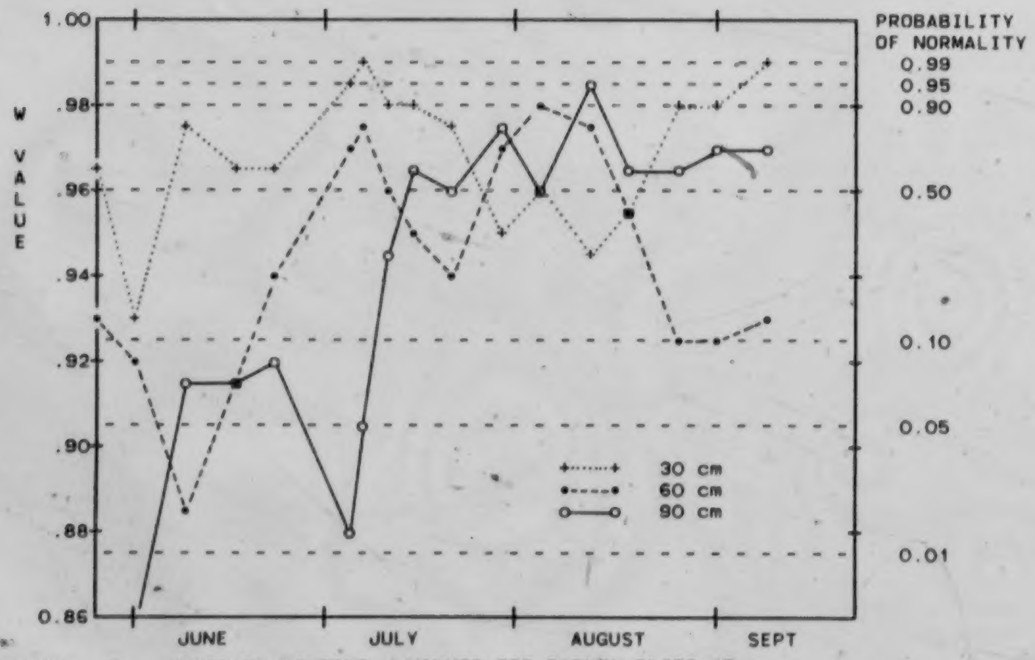

FIGURE 26. MOISTURE CONTENT W VALUES FOR BARLEY PLOTS AT 30, 60, AND 90 CM

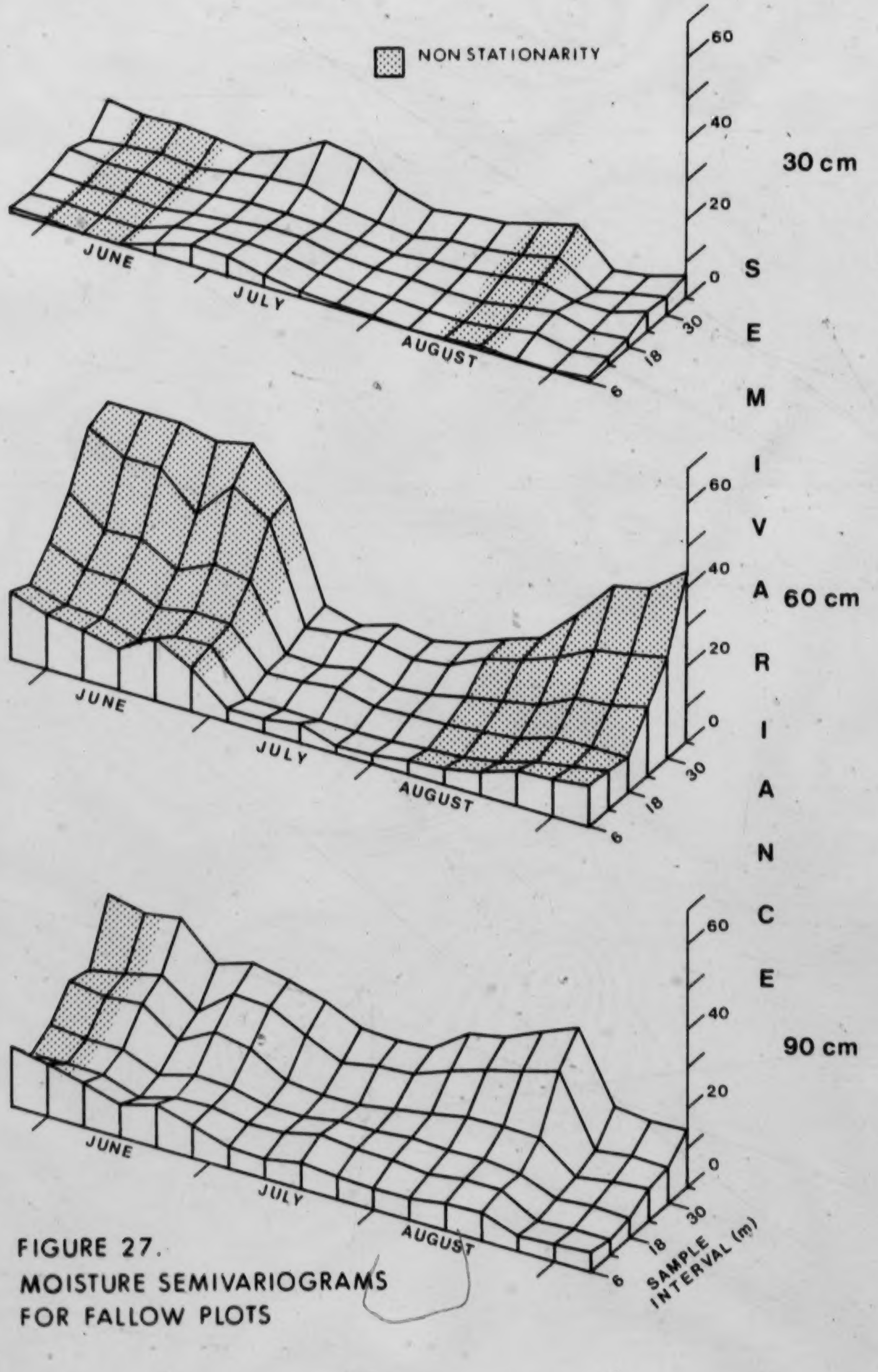

FIGURE 27.
MOISTURE SEMIVARIOGRAMS
FOR FALLOW PLOTS

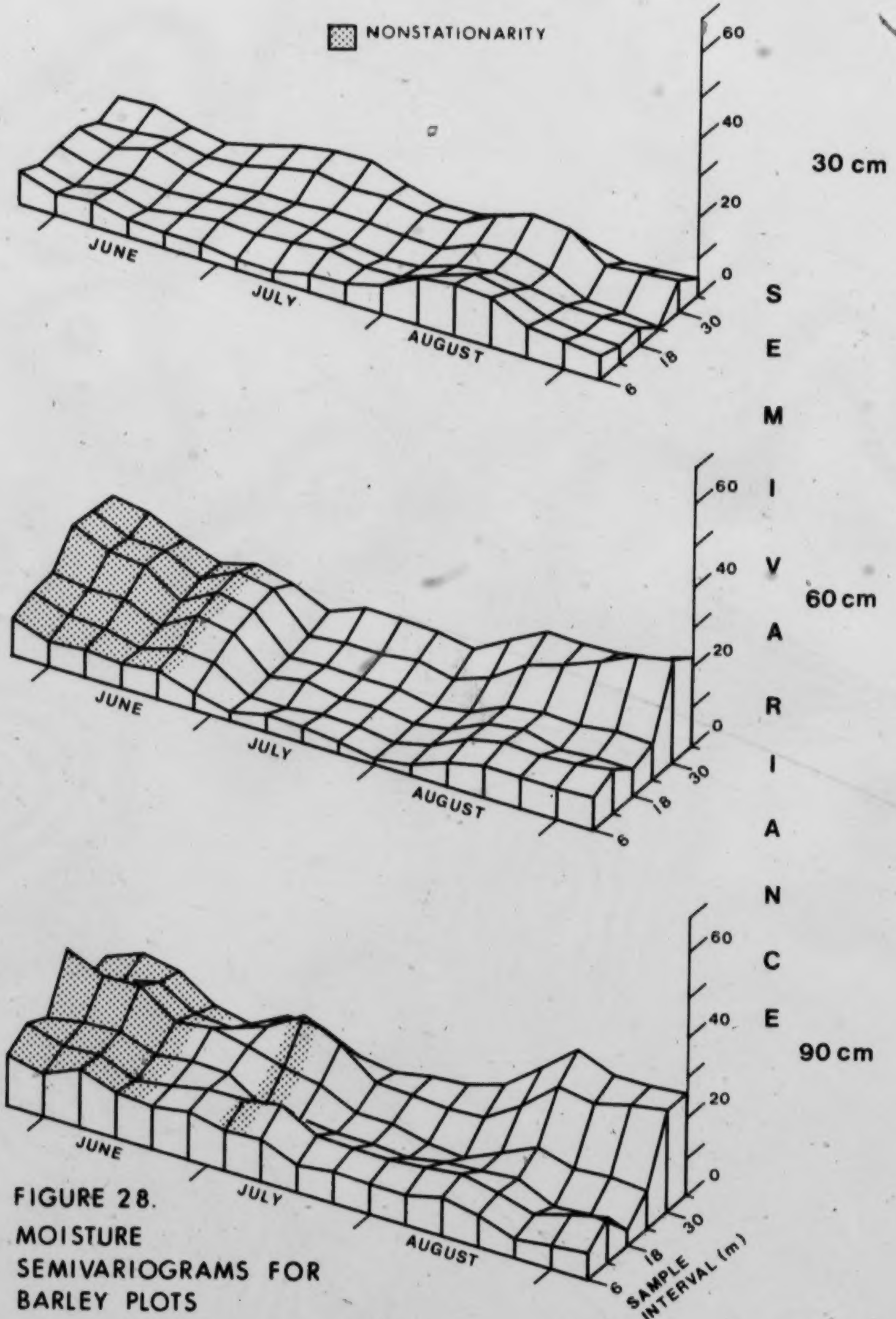

FIGURE 28.
MOISTURE
SEMIVARIOGRAMS FOR
BARLEY PLOTS

only from semivariances calculated for the north-south vectors (Table 13). Nonstationarity resulting from significant anormality is marked with a stippled area.

In general the semivariograms for stationary conditions are indicative of spatial independence, that is, moisture contents measured at neighbouring sites were not any more similar than those further away. Although the semivariance does slightly increase with distance the increase is not sufficiently strong to be indicative of spatial dependence. Assuming that spatial dependence does exist but it is a matter of scale then the present sample interval of 6.1 m is too large or/and it is too small.

Other trends of note from the moisture content semivariograms displayed in Figures 27 to 29 are:

1. smaller semivariances for stationary conditions at the 30 and 60 cm depths than the 90 cm depth;
2. a decrease in semivariance and in slope of semivariogram for wet conditions during early July, except for the fallow plots at 30 cm; and
3. indications of spatial dependence occurring for moisture contents at 90 cm during August, but with discontinuity at the origin and no apparent sill values (Type S3, Figure 20c, Page no. 106)

Interpretations of the effects of barley as compared to fallow conditions upon the moisture content semivariances could not be evaluated due to nonstationarity and textural variability within the site.

4.5.3 SOIL WATER POTENTIAL

4.5.3.1 Precision and Distribution

The mean, coefficient of variation, and sample sizes for precisions of 10%, for tensiometers at 30 and 60 cm depth are listed in Table 17. The variability of the 30 and 60 cm tensiometer readings were greater than the CV's for other soil properties. Generally more than 30 samples were required to achieve a precision of ±10%. The readings for the 60 cm tensiometers were more variable than the 30 cm readings. Barley plot readings were more variable than those for fallow. There does not appear to be any apparent trend of increasing variability with decreasing potential.

Although the sample sizes were not large enough for any valid statistical comparisons using the skew and kurtosis parameters, the skew values, with the exception of the 60 cm depth for barley plots, were generally low (below 1.0). None of the *W* values were significantly anormal nor normal at a probabilty level of ≤ 0.05.

Logarithmic transformations of the absolute potential values, expressed in cm of water (*pF*), resulted in decreased variability as indicated by the CV values and the sample size requirements, but did not always result in decreased skew or *W* values.

TABLE 17. Statistics for the 30 and 60 cm Tensiometers

DATE	N	MEAN (-kPa)	CV	D10	SKEW	*W*
30 cm depth						
July 14	17	8.5	18.6	15	0.00	0.966
log		1.91	5.1	1	-0.56	0.940
July 21						
Barley	7	11.3	36.1	67	-0.01	0.920
log		1.97	8.0	4	0.12	0.902
Fallow	9	9.3	23.5	26	-0.56	0.963
log		1.96	5.8	2	-0.79	0.910
July 28						
Barley	7	22.7	25.7	32	-0.45	0.914
log		2.34	5.2	2	-0.53	0.889
Fallow	9	15.1	11.7	7	-0.37	0.933
log		2.17	2.4	1	0.29	0.916
August 11						
Barley	9	79.5	8.9	4	0.61	0.904
log		2.90	1.3	1	0.09	0.916
Fallow	9	44.3	22.2	22	-0.19	0.956
log		2.64	3.9	1	-0.48	0.931
60 cm depth						
July 14	18	3.2	38.6	63	0.34	0.956
log		1.48	12.0	6	-0.23	0.963
July 21						
Barley	9	4.8	41.5	81	0.96	0.902
log		1.65	10.2	5	0.31	0.972
Fallow	9	4.6	36.9	64	-0.13	0.870
log		1.63	10.8	5	-0.29	0.876
July 28						
Barley	9	6.6	48.8	112	1.36	0.846
log		1.81	9.7	5	0.50	0.961
Fallow	9	4.6	31.5	47	0.24	0.976
log		1.64	8.3	4	-0.17	0.980
August 11						
Barley	8	41.5	46.5	106	-0.17	0.942
log		2.56	10.3	6	-0.87	0.875
Fallow	9	6.7	58.8	163	0.80	0.897
log		1.76	14.7	11	-0.27	0.954

Log values are derived from $\log_{10}$ cm H_2O

4.5.4 CHANGES IN SOIL MOISTURE

4.5.4.1 Precision

The CV for total changes in soil moisture, ΔS, within the top 1 m of the soil profile, is graphed against time in Figure 29. Generally the variability of changes in soil moisture were greater than any other soil property especially during May to late June and late August when the CV values were over 80%. CV values ranged from a low of 18% for the barley plots in August to values greater than 1000% in September. Generally CV's were very large (>100%) when ΔS was very small (late May to late June and September), large (40 to 80%) when ΔS was increasing due to precipitation, and low (<40%) during conditions of high *Et*.

Required sample sizes varied proportionally with the CV's between 20 and 2000 for a precision of ±10%.

4.5.4.2 Distribution

Skew values and *W* values for total change in soil moisture within the top 1 m of the soil profile are plotted against time in Figures 30 and 31. Significantly anormal distributions occurred in June with a negative skew and in early July, during large increases in ΔS due to rainfall, with a positive skew. During late July and August when *Et* was large, the ΔS distributions as indicated by the skew and *W* values were normal.

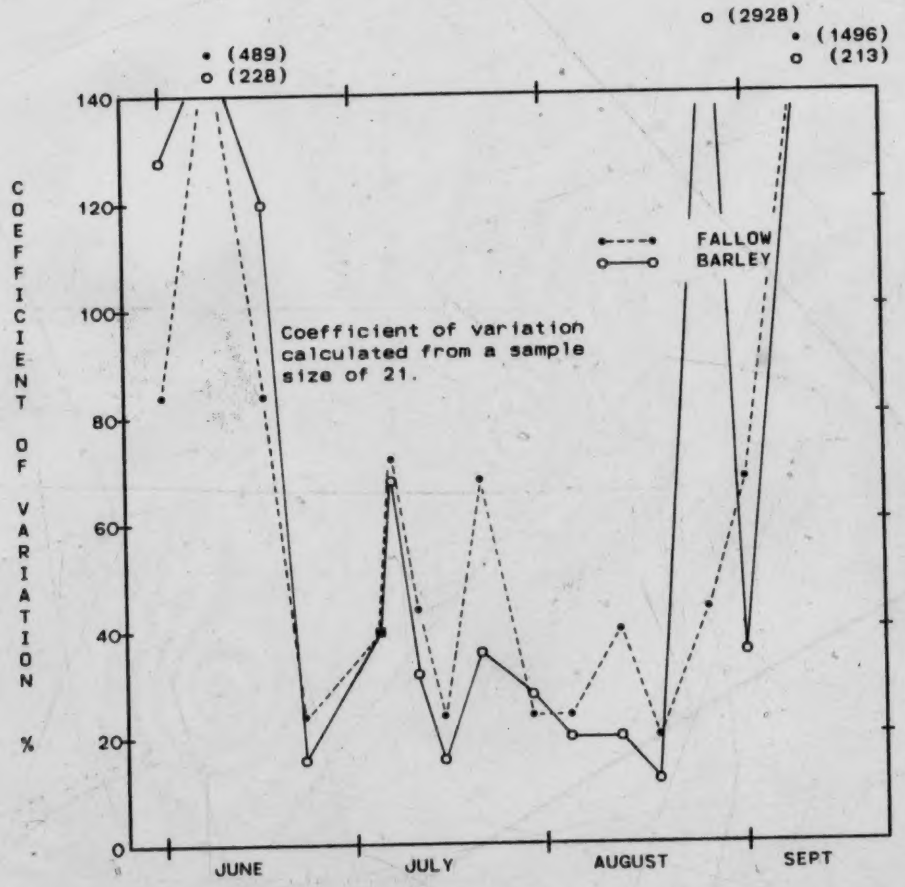

FIGURE 29. COEFFICIENT OF VARIATION FOR CHANGE IN PROFILE MOISTURE STORAGE FOR FALLOW AND BARLEY PLOTS

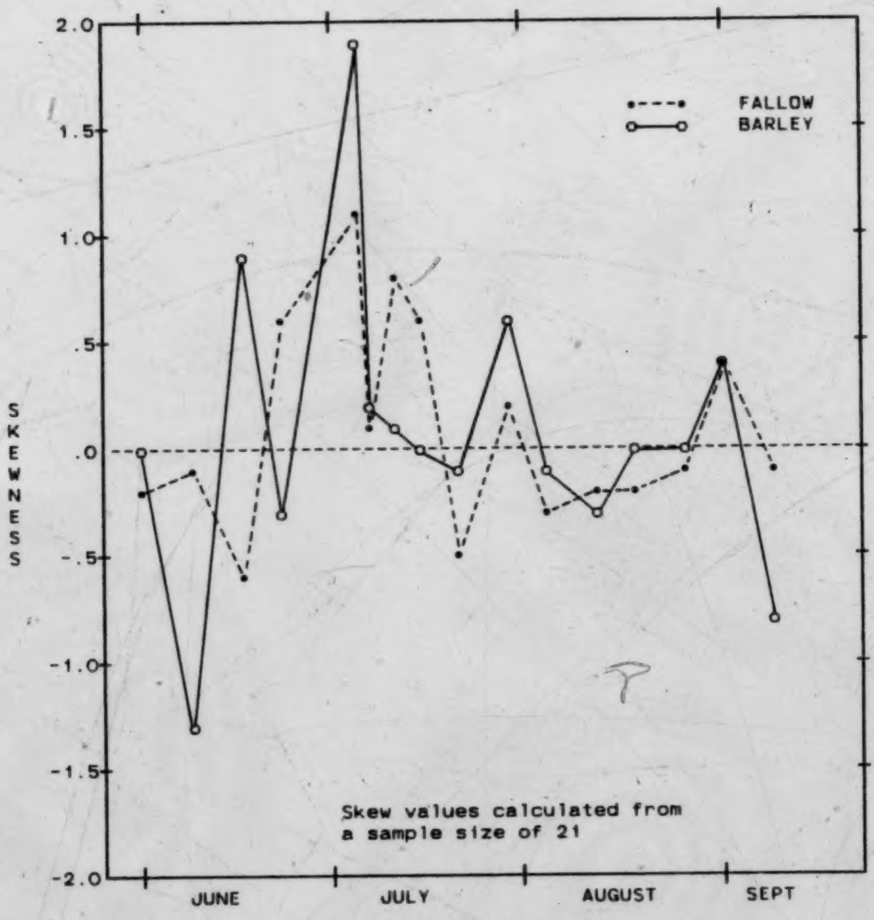

FIGURE 30. SKEWNESS OF CHANGES IN PROFILE MOISTURE STORAGE FOR FALLOW AND BARLEY PLOTS

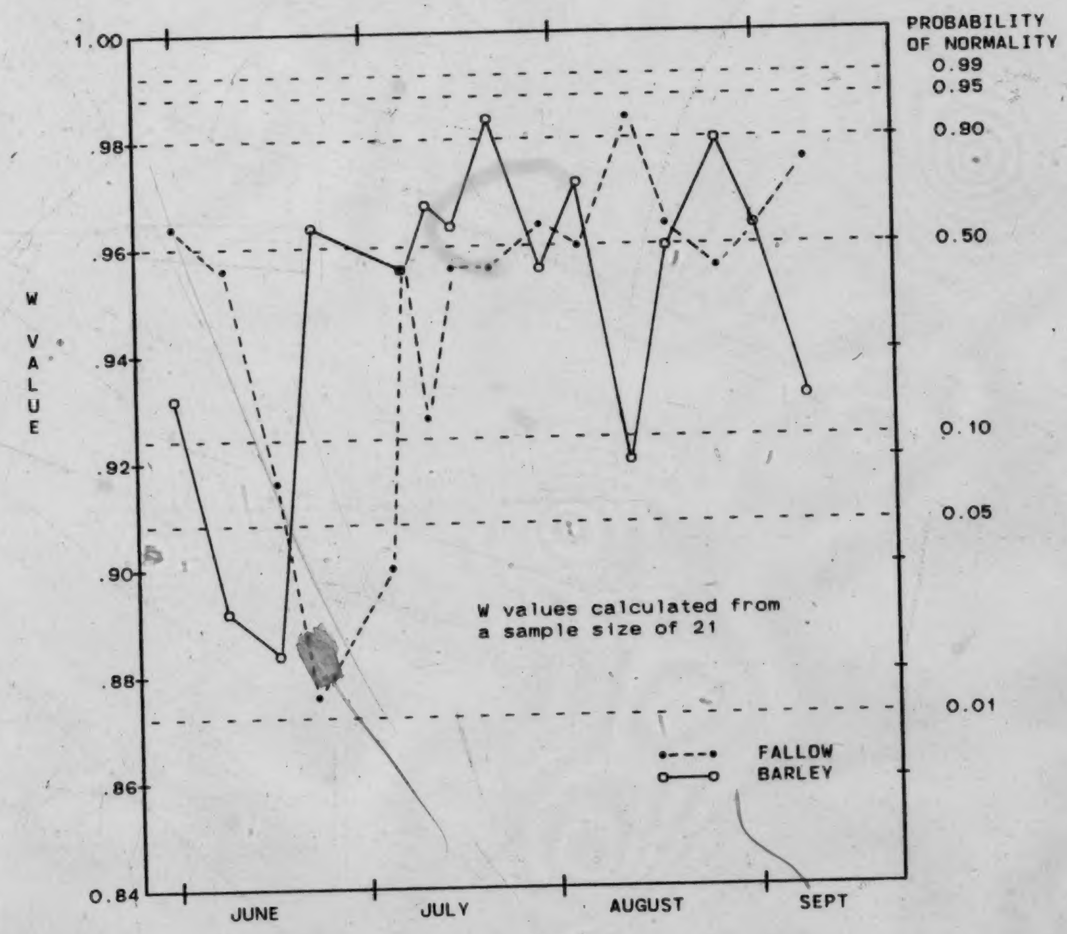

FIGURE 31. W VALUES FOR CHANGES IN SOIL PROFILE MOISTURE FOR FALLOW AND BARLEY PLOTS

4.5.4.3 Semivariance

The semivariances varied greatly depending upon the vector and also among different time periods. The semivariograms were characterized by large nugget values, ranging from 10 to over 400. No spatial dependence within these sampling distances are indicated with any consistency.

5. SUMMARY AND CONCLUSIONS

The water balance was determined for fallow and barley moisture regimes from soil moisture measurements in conjunction with precipitation measurements and evapotranspiration calculations. Soil moisture was measured with a neutron probe and evapotranspiration was calculated using Penman's equation. Systematic location of sample points enabled assessment of the spatial variability of soil moisture.

Drainage and evapotranspiration were quantified using a combination of calculated *Et* by the Penman equation for conditions of positive ΔS and by the field capacity (fallow and barley) and the gradient methods (barley only) during conditions of negative ΔS. Drainage from May 26 to September 8, 1983, calculated using the field capacity method as 94 mm for both the fallow and the barley plots. Evapotranspiration was estimated tobe 207 and 254 mm for the fallow and barley plots. Results using the gradient method differed little from the field capacity method: drainage and evapotranspiration for the barley plots were respectively 88 and 260 mm. The slight difference between the two methods is partially masked by the occurrence of about 90% of the drainage during the wet conditions. The Penman method was used between June 16 and July 7 when 224 mm or 80% of the season's precipitation occurred. Drainage calculated using the field capacity and the gradient methods (July 7 to September 8) was estimated for the barley plots at 12 and 6 mm respectively. The largest potential source of error lies in the estimation of actual evaporation by the application of a coefficient (0.8) to *PET* calculated by the Penman method. Although no measurement of the actual evaporation was available, an error as much as ±10% might be expected. The moisture deficit for the entire study period was 17 and 64 mm for the fallow and the barley plots respectively.

Due to the probable occurrence of upward flow from a shallow water table to the root zone, a precise estimation of evapotranspiration was not possible. The evapotranspiration calculated here is undoubtedly low due to upward flow of water. The literature indicates that a water table at a depth of 1 to 2 m in the presence of an actively growing crop will result in upward flow and could contribute 30% or greater of evapotranspiration. A contribution of 30% would account for all the water that was attributed to drainage and consequently the drop in the water table would be entirely attributable to upward flow.

The only moisture input to the site from May 26 to September 8 was precipitation. The water table did not exist before June 23, the onset of heavy rains. Runoff and interflow onto the site was likely nil due to the presence of a gentle slope (about 1%)at the high portion of the site and the surrounding grass strips. Within site, runoff was probably greater, as indicated by micro-rills, and could have accumulated in the lower portion of the site where the slope was steeper (3%). Interflow could also have contributed to moisture accumulation and possibly outflow from the northwest corner of the site due to the steeper topography and the slope in the water table. A slight increase in soil moisture was evident in the northwest corner, but was confounded by the water table being much closer to the surface.

The spatial variability of soil moisture and other soil properties was determined by calculating the precision of the measurements, the normality of the distribution, and the spatial dependence of the sample intervals.

The number of samples required to measure soil moisture with a precision of ±5% for a probability of 95% varied from 28 for the barley plots at 30 cm in the beginning of August to two samples at 90 cm in mid-July for the barley plots. During wet conditions in the first week of July the 30 cm moisture content in the barley plots required only four samples. Generally soil moisture precision increased as the moisture content increased. Soil properties measured at this site are ranked below in terms of relative precision from highest to lowest:

i) soil moisture (near saturation)
ii) bulk density
iii) soil moisture (dry)
iv) moisture -33 kPa
v) moisture -1500 kPa
vi) sand,silt,clay
vii) soil suction
viii) ΔS

Most of these properties, including soil moisture, had the lowest precision at depths below 60 cm. This was due to the presence of discontinuous sand lenses. To determine ΔS within ±5% of the mean required between 20 and 2000 samples depending upon the value of ΔS. Generally the precision of ΔS increased as the absolute value of ΔS became larger, especially for losses due to *Et*.

Soil moisture became more anormally distributed (higher skew values) as soil moisture decreased. During near saturated conditions, skew values near zero indicated normal distributions, whereas for drier conditions the moisture content, especially for the deeper depths, was significantly negatively skewed. Soil texture, and moisture contents at 33 and 1500 kPa were often significantly anormal at the deeper depths also. Changes in soil moisture were more normal for conditions of high evapotranspiration than for conditions of high precipitation and drainage.

Semivariograms of all soil properties indicated that neighbouringsamples were generally spatially independent. There was a lack of spatial dependence at the sampling interval used (6.1 m) over the distance sampled (43.1 m). The large nugget values relative to the semivariances at greater sample intervals indicated that most of the variance occurred at distances less than 6.1 m.

The lack of any spatial dependence at the measured distances could be in part due to the high variabilities encountered. These high variabilities likely resulted in the assumption of stationarity not being met.

Proper statistical evaluation of soil moisture between two different vegetative regimes requires a certain number of samples spaced a certain distance apart. The first requirement to accomplish this is to establish the desired precision. Research work may require a greater precision, (±5%) than applications or survey studies (±10%). Site conditions for this study dictate that upwards of 28 neutron moisture measurements for one depth are required to achieve a precision of ±5% within an 0.1 ha site. During near saturated conditions this sample size for the same depth reduced to 4. The spatial independence of the study site indicated that the samples may be spaced 6 to 37 m apart and that a random spacing would offer the same efficiency in precision as would a systematic spacing. If spatial dependence does occur at distances smaller than 6 m or larger than 37 m, then systematic spacing would result in increased precision over random spacing.

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APPENDIX A

PROFILE DESCRIPTION OF SOIL PROFILE AT ELLERSLIE RESEARCH STATION. (Crown and Greenlee, 1978)

Soil: Eluviated Black Chernozem (Typic Argialboll)

Location: Ellerslie Research Station, Edmonton, Alberta, NE 1/4 Sec 24, Tp 51, R 25, W 4, (Figure 1).

Parent Material: Lacustrine with interbedded till.

Landform and Site Position: Undulating, upper slope position, very gently sloping to the East.

Soil Drainage: Well drained.

Present Landuse: Recently abandoned cropland, bounded by active cropland.

Vegetation: Grasses and some common weeds.

HORIZON	DEPTH (cm)	DESCRIPTION
Ah	0-30	Black (10YR 2.5/1 m) silt loam; weak to moderate, medium subangular blocky breaking to moderate to strong, fine granular; friable; abundant, fine random roots; many, fine, random pores; no clay films; no carbonates; no coarse fragments; abrupt, wavy boundary; 23 to 34 cm thick; slightly acid to neutral reaction.
Ae	30-37	Brown (10YR 5/3 m) silt loam; weak to moderate, fine platy; friable; plentiful, fine, vertical roots; many, fine random pores no clay films; no carbonates; no coarse fragments; clear, broken boundary; 1 to 10 cm thick; slightly acid reaction.
Bt1	37-87	Brown (10YR 4/3 m) loam; moderate, medium to coarse prismatic; firm; plentiful, fine vertical roots; common, very fine, vertical pores; many, moderately thick, dark grayish brown (10YR 4/2) clay films in many voids and on many ped surfaces; estimated 10% gravelly coarse fragments; gradual, wavy boundary; 43 to 61 cm thick; slightly acid reaction.
Bt2	87-113	Dark grayish brown (10YR 4/2 m) clay loam; moderate, coarse prismatic; very firm;

		few, very fine, vertical roots; common, very fine, vertical pores; many, moderately thin, dark grayish brown (10YR 4/2) clay films in many voids and on many vertical ped surfaces; no carbonates; estimated 10% gravelly and 2% cobbly coarse fragments; abrupt, wavy boundary; 22 to 33 cm thick; neutral reaction.
Ccag	113+	Yellowish brown (10YR 5/3.5 m) silt loam; few, medium, prominent, dark reddish brown (2.5YR 3/4) and strong brown (7.5YR 5/6) mottles; massive; very firm; no roots; few, very fine, random pores; no clay films; moderately calcareous; secondary carbonates common, medium light gray (10YR 7/2) vertical streaks and irregular spots, very friable; estimated 5% gravelly coarse fragments; moderately alkaline reaction.